

The final publication is available at Elsevier via <http://dx.doi.org/10.1016/j.lithos.2017.01.009> © 2017. This manuscript version is made available under the CC-BY-NC-ND 4.0 license <https://creativecommons.org/licenses/by-nc-nd/4.0/>

Behaviour of apatite during partial melting of metapelites and consequences for prograde suprasolidus monazite growth

Chris Yakymchuk

Department of Earth and Environmental Sciences
University of Waterloo
Waterloo, Ontario, Canada, N2L 3G1
email: cyakymchuk@uwaterloo.ca

Abstract

The suprasolidus behaviour of apatite and monazite are examined for an average metapelite composition using phase equilibria modelling coupled with solubility equations of these minerals. Both closed- and open-system scenarios are considered. Partial melting above the solidus requires apatite and monazite breakdown in order to saturate the anatectic melt in phosphorus and the light rare earth elements. In general, melt loss is predicted to increase the stability of apatite and monazite at high temperature. Most apatite is predicted to survive up to ultrahigh temperature conditions except for rocks with low bulk phosphorus concentrations. By contrast, most monazite is expected to be consumed by UHT conditions. Thorium substitution in monazite is expected to increase the stability of monazite to higher temperatures. The presence of LREE-rich apatite decreases the stability of monazite above the solidus, but the breakdown of this apatite during anatexis may generate prograde monazite at the apatite–melt interface in local pockets of melt oversaturation. However, prograde suprasolidus monazite along grain boundaries is expected to be consumed during further partial melting or during melt homogenization when an interconnected melt network develops. Anatectic melts are predicted to be saturated with respect to apatite except at UHT conditions and for rocks with low initial P_2O_5 bulk concentrations.

Keywords: Accessory Minerals, Migmatite, Granulite, Partial Melting, Apatite, Monazite

1. Introduction

30 Apatite and monazite are ubiquitous minerals in aluminous metamorphic rocks and are the
31 dominant repository for phosphorus and the light rare earth elements (LREE) in migmatites and
32 granulites (e.g. Bea et al., 1994; Bea, 1996a, 1996b; Spear and Pyle, 2002; Prowatke and Klemme,
33 2006). While apatite is commonly used for geochronology of low-temperature processes (e.g. Farley and
34 Stockli, 2002), monazite is frequently used to date portions of suprasolidus P – T paths in high-
35 temperature metamorphic rocks (e.g. Parrish, 1990; Lederer et al., 2013; Taylor et al., 2016). Both
36 minerals are expected to break down in suprasolidus metamorphic rocks during heating and anatexis in
37 order to saturate the anatectic melt in phosphorus and the LREE (e.g. Watson, 1979; Watson and Harrison,
38 1983; Rapp and Watson, 1986; Montel, 1986, 1993; Rapp et al., 1987; Pichavant et al., 1992; Wolf and
39 London, 1994; London et al., 1999; Stepanov et al., 2012; Duc-Tin and Keppler, 2015).

40 Theoretical modelling of monazite behaviour in suprasolidus metamorphic rocks suggests that
41 prograde monazite growth above the solidus is unlikely and that monazite ages record cooling and
42 crystallization from anatectic melt (e.g. Kelsey et al., 2008; Spear and Pyle, 2010; Yakymchuk and
43 Brown, 2014b). This contrasts with several studies that convincingly show that monazite can record
44 prograde ages associated with heating and partial melting (e.g. Hermann and Rubatto, 2003; Dumond et
45 al., 2015; Hacker et al., 2015; Johnson et al., 2015; Blereau et al., 2016). Therefore, the mechanisms that
46 generate prograde monazite in suprasolidus metamorphic rocks are unclear. Furthermore, the behaviour
47 of LREE-rich apatite during suprasolidus metamorphism has not yet been quantified using phase
48 equilibria modelling, but the breakdown of this apatite and growth of monazite has been documented in
49 experimental studies (e.g. Wolf and London, 1994) and has been proposed as a possible mechanism to
50 produce prograde suprasolidus monazite (e.g. Johnson et al., 2015). However, the coupling of LREE-
51 rich apatite to monazite behaviour during anatexis has yet to be evaluated using quantitative phase
52 equilibria modelling because previous studies did not include phosphorus in their models (e.g. Kelsey et
53 al., 2008; Yakymchuk and Brown, 2014b).

54 In this contribution, I first review the controlling factors for apatite and monazite dissolution in
55 anatectic melt. Then I combine phase equilibria modelling of an anatectic metapelite with the
56 experimental results of apatite and monazite dissolution to quantify the coupled behaviour of apatite and
57 monazite during anatexis as well as evaluate if the breakdown of LREE-rich apatite can trigger prograde
58 monazite growth above the solidus. Both closed-system and open-system (e.g. melt loss) scenarios are
59 considered. The consequences for phosphorus saturation of anatectic melts and granites with respect to
60 apatite are also discussed.

61

62 **2. Suprasolidus behaviour of apatite and monazite**

63 Apatite and monazite are the main carriers of phosphorus in suprasolidus metamorphic rocks and
64 their breakdown is required to maintain phosphorus saturation of melt during anatexis. In addition,
65 monazite is the primary repository of LREE in most metasedimentary rocks (Bea, 1996b) and it is the
66 main contributor of the LREE to anatectic melt. Monazite breakdown will also liberate phosphorus—an
67 essential structural constituent of monazite (~29 wt.% P_2O_5)—into the anatectic melt, although the
68 contribution is expected to be negligible compared with any phosphorus released from apatite
69 breakdown (e.g. London et al., 1999; García-Arias et al., 2012). Apatite can also contain significant
70 concentrations of the LREE (e.g. Bea et al., 1994) and its breakdown may release LREE into the melt.
71 Therefore, the breakdown and growth of apatite and monazite in suprasolidus systems depend on each
72 other; the breakdown of one may lead to oversaturation of the melt with respect to the other.

73 The breakdown of apatite and monazite into melt is controlled by chemical and physical factors,
74 including: temperature, pressure, melt chemistry and kinetics (Watson, 1979; Watson and Green, 1981;
75 Watson and Harrison, 1983; Rapp and Watson, 1986; Piccoli and Candela, 2002; Stepanov et al., 2012).
76 Several experimental studies have examined these factors for peraluminous melt compositions, which
77 are applicable to metapelites and metagreywackes (e.g. Harrison and Watson, 1984; Bea et al., 1992;
78 Pichavant et al., 1992; Wolf and London, 1994; Toplis and Dingwell, 1996; London et al., 1999).

79 Harrison and Watson (1984) used partial melting experiments to show that apatite solubility is
80 related to temperature and the SiO₂ concentration of melt by the following relationship:

$$81 \quad \ln D_P^{apatite/melt} = \left[\frac{(8400 + ((SiO_2 - 0.5)2.64 \times 10^4))}{T} \right] - [3.1 + 12.4(SiO_2 - 0.5)] \quad (1)$$

82 where SiO₂ is the weight fraction in the melt and T is in Kelvin. This relationship is predicted to be valid
83 for melt with SiO₂ concentrations between 45% and 75%, for 0% to 10% wt.% H₂O and for the range of
84 pressures expected in the crust (Harrison and Watson, 1984). Pichavant et al. (1992) conducted
85 reconnaissance experiments and they demonstrated that apatite solubility increases for peraluminous
86 melt compositions compared with the metaluminous and peralkaline melts used by Harrison and Watson
87 (1984). Building on the work of these previous studies, Wolf and London (1994) derived a simple
88 equation that relates the aluminum saturation index (ASI: molar Al₂O₃ / [Na₂O + K₂O + CaO]) of the
89 melt to the concentration of P₂O₅ in melt that is in chemical equilibrium with apatite:

$$90 \quad P_2O_5 (wt.%) = -3.4 + 3.1 \times ASI \quad (2)$$

91 However, their experiments were restricted to low pressure (2 kbar) and peraluminous melt
92 compositions (ASI > 1.1), which may be applicable for partial melting of aluminous metasedimentary
93 rocks but not for intermediate to basic rocks.

94 The solubility of monazite in granitic melts has been investigated in several experimental studies
95 (Montel, 1986; Rapp and Watson, 1986; Rapp et al., 1987; Montel, 1993; Stepanov et al., 2012; Duc-Tin
96 and Keppler, 2015). Stepanov et al. (2012) combined their experimental results with previous work and
97 derived the following equation for monazite solubility:

$$98 \quad \ln \Sigma LREE = 16.16(\pm 0.3) + 0.23(\pm 0.07)\sqrt{H_2O} - \frac{11494(\pm 410)}{T} - 19.4(\pm 4)P/T + \ln X_{mnz}^{LREE} \quad (3)$$

99 where the amount of H₂O in melt is in wt.%, T is in Kelvin, P is in kbar, $\Sigma LREE$ is the sum of La–Sm
100 (in ppm) in the melt and X_{mnz}^{LREE} is the molar ratio of the LREE to the sum of all cations in monazite. In
101 general, monazite solubility is higher for increasing T and the H₂O concentration of the melt and
102 monazite solubility decreases with increasing pressure.

103

104 **3. Methodology**

105 *3.1. Phase equilibria modelling*

106 An average amphibolite-facies metapelite (Ague, 1991; Table 1) is used to model the
107 suprasolidus phase relations, melt compositions and behaviour of apatite and monazite during partial
108 melting. This composition is chosen because it is generally the most fertile rock type in metasedimentary
109 sequences and metapelites usually contain low-variance assemblages that are particularly useful for
110 metamorphic studies. Calculations were performed using THERMOCALC v.3.40 (Powell and Holland,
111 1988) and the internally consistent dataset of Holland and Powell (2011). Modelling was undertaken in
112 the MnO–Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂–Fe₂O₃ (MnNCKFMASHTO) chemical
113 system (e.g. White et al., 2014a, b). The activity–composition models are from White et al. (2014a).
114 Phases modelled as pure end-members include quartz, rutile, aqueous fluid (H₂O), kyanite and
115 sillimanite. Mineral abbreviations are from Holland and Powell (2011).

116 Suprasolidus rocks are generally not expected to have excess free H₂O due to the low
117 permeability of the deep crust and any free H₂O is likely partitioned into anatectic melt (White and
118 Powell, 2002; White et al., 2005). The solubility of H₂O in anatectic melt increases with pressure along
119 the wet solidus. For the phase equilibria modelling presented here, the amount of H₂O in the bulk
120 composition is adjusted so that the melt is just saturated with H₂O at solidus where the modelled
121 prograde path crosses the solidus. If the prograde path crossed the wet solidus at lower or higher
122 pressures, the quantity of melt produced will be slightly overestimated and underestimated, respectively.
123 Also, the melt at the wet solidus above this pressure will be slightly oversaturated in H₂O.

124 The average metapelite composition saturated with H₂O at the solidus at 9 kbar was used to
125 model closed-system behaviour. This does not account for melt loss, which is a path-dependent process
126 (e.g. Mayne et al., 2016; Guevara and Caddick, 2016) and this is discussed next. However, the closed-

127 system model is used as a baseline to assess the sensitivity of the modelling to different variables related
128 to the behaviour of apatite and monazite.

129

130 *3.2 Open-system behaviour*

131 The preservation of high-grade metamorphic assemblages in migmatites and granulites requires
132 the loss of anatectic melt (e.g. Spear et al., 1999; White and Powell, 2002). Melt loss during anatexis has
133 important implications for the behaviour of major and accessory minerals during metamorphism (e.g.
134 Yakymchuk and Brown, 2014b) as well as the concentration of the LREE and phosphorus in the
135 residuum (e.g. Rapp et al., 1987). The effects of melt loss on the behaviour of apatite and monazite are
136 modelled for two schematic P – T paths that involve isobaric heating from the solidus up to 950°C at 9
137 kbar and 6 kbar. The higher pressure isobaric path was chosen to approximate a typical prograde path
138 for collisional orogenesis (e.g. Clark et al., 2011) and the lower pressure path to model the reaction
139 sequence experienced by high-pressure–low-temperature terranes (e.g. Kelsey and Hand, 2015). For the
140 isobaric heating path at 6 kbar, the starting bulk composition is adjusted so that the melt is just saturated
141 with H₂O at solidus at this pressure.

142 Melt loss is modelled following the methodology of Yakymchuk and Brown (2014a, b) whereby
143 6 mol.% of melt is extracted out of the system when the total amount reaches 7 mol.% (approximately
144 equivalent to 7 vol.% on a one-oxide basis). The modelled 7 vol.% threshold approximates the melt
145 connectivity threshold of Rosenberg and Handy (2005). The extraction of 6 mol.% melt leaves 1 mol.%
146 melt in the system. The extracted melt has a major element composition determined from the phase
147 equilibria modelling and phosphorus and LREE concentrations determined from the saturation models
148 above (equations 1–3) as long as apatite and monazite are present in the system to buffer the
149 concentrations. After each melt loss event, the melt-depleted composition is used to model the next
150 segment of the isobaric heating path up to the next melt loss event. This process is repeated until the end
151 of each isobaric heating path is reached at 950°C.

152

153 3.3. Monazite and apatite behaviour

154 The amounts of apatite and monazite dissolution and growth are modelled by similar methods to
155 those reported in Kelsey et al. (2008). The limitations and assumptions of this approach are discussed
156 later. First, the saturation concentrations of the melt in ppm are calculated as follows. The major element
157 concentration of the anatectic melt at P – T is calculated using THERMOCALC every 1°C and 0.1 kbar
158 across the suprasolidus portion of the pseudosection shown in Figure 1. This information is combined
159 with experimentally determined solubility equations for apatite of Harrison and Watson (1984; HW84)
160 as well as Wolf and London (1994; WL94). Both equations are used because they are both applicable to
161 peraluminous melts, but they yield different results. The solubility equation for monazite is from
162 Stepanov et al. (2012). The solubility equations are combined with assumed stoichiometric
163 concentrations of phosphorus in apatite (41 wt.% P_2O_5 ; the average of apatite compositions reported in
164 Webster and Piccoli, 2015), LREE in monazite (566,794 ppm; Kelsey et al., 2008) and phosphorus in
165 monazite (29 wt.% P_2O_5). The result is the saturation concentration of phosphorus (P_{Sat}) and the LREE
166 ($LREE_{Sat}$) in the anatectic melt at each P – T point. These ‘initial saturation concentrations’ are
167 independent of the proportion of anatectic melt and the bulk concentrations of phosphorus and LREE.

168 Three bulk rock values of P_2O_5 (0.1, 0.3 and 0.5 wt.%) are used that represent the typical range
169 of measured compositions in pelites (e.g. Ague, 1991; London et al., 1999). A bulk LREE concentration
170 of 150 ppm is used that is representative of pelites and metasedimentary migmatites (e.g. Taylor and
171 McLennan, 1985; Yakymchuk and Brown, 2014b).

172 At this point the methodology departs from that of Kelsey et al. (2008) in order to couple the
173 behaviour of monazite to that of apatite. All of the phosphorus and LREE in the bulk rock are assumed
174 to reside only in apatite, monazite and melt. No phosphorus substitution into major minerals is
175 considered; the possible effects of this are discussed later. The amounts of these elements are then
176 allocated from the bulk concentrations of phosphorus and LREE to form apatite and monazite based on

three values: (1) the bulk concentration of LREE in ppm ($LREE_{Bulk}$), (2) the bulk concentration of phosphorus in ppm (P_{Bulk}), and (3) the concentration of LREE in apatite in ppm (Ap_{LREE}). The concentration of LREE in apatite will vary with melt composition (Prowatke and Klemme, 2006) due to partitioning of LREE between apatite and melt and the concentration of LREE in apatite can be determined by:

$$Ap_{LREE} = D_{LREE}^{Ap/melt} \times LREE_{melt} \quad (4)$$

where $D_{LREE}^{Ap/melt}$ is the partition coefficient of LREE between apatite and melt and $LREE_{melt}$ is equal to $LREE_{Sat}$ when monazite is present in the system. Here, I use a $D_{Ap/melt}^{LREE}$ value of 10, which is an average of the experimental results reported in Prowatke and Klemme (2006) for La partitioning between apatite and melt for melts with concentrations of >45 wt.% SiO₂. The amounts of phosphorus in apatite (Ap_P) and monazite (Mnz_P) and LREE in monazite (Mnz_{LREE}) are fixed to the stoichiometric values stated above.

The bulk concentration of phosphorus and LREE is the sum of these elements allocated to each mineral:

$$P_{Bulk} = P_{Ap} + P_{Mnz} \quad (5)$$

$$LREE_{Bulk} = LREE_{Ap} + LREE_{Mnz} \quad (6)$$

where P_{Ap} and P_{Mnz} are the amounts of phosphorus allocated to form apatite and monazite, respectively and $LREE_{Ap}$ and $LREE_{Mnz}$ are the amount of LREE allocated to form apatite and monazite, respectively.

Starting with the assumption that all LREE in the rock resides in monazite ($LREE_{Bulk} = LREE_{Mnz}$), and that the ratio of $\frac{Mnz_P}{Mnz_{LREE}}$ is fixed, the amount of phosphorus needed to form stoichiometric monazite is calculated as follows:

$$P_{mnz} = \frac{Mnz_P}{Mnz_{LREE}} \times LREE_{Mnz} \quad (7)$$

The remaining amount of phosphorus that is allocated to form apatite is then calculated by rearranging equation (5):

$$P_{Ap} = P_{bulk} - P_{mnz} \quad (8)$$

Using this amount of phosphorus available to form apatite, the amount of LREE allocated to make apatite with the concentration of LREE (Ap_{LREE}) determined by partitioning between apatite and melt (equation 4) is then calculated:

$$LREE_{Ap} = \frac{P_{Ap}}{\frac{Ap_P}{Ap_{LREE}}} \quad (9)$$

The amount of LREE remaining to form monazite is then calculated by rearranging equation (6):

$$LREE_{Mnz} = LREE_{Bulk} - LREE_{Ap} \quad (10)$$

Here, the revised amount of LREE allocated to monazite ($LREE_{Mnz}$) is inserted back into equation (7) and this series of equations (7–10) is solved iteratively to determine the amounts of phosphorus and LREE allocated to apatite and monazite, respectively. These values are equivalent to the ‘effective’ concentrations of Kelsey et al. (2008) and Yakymchuk and Brown (2014b) and are independent of the amount of melt in the system.

The saturation concentrations of phosphorus (P_{Sat}) and the LREE ($LREE_{Sat}$) calculated from the solubility equations of apatite and monazite (Harrison and Watson, 1984; Wolf and London, 1994; Stepanov, al. 2012) are now divided by the allocated values of phosphorus to apatite (P_{Ap}) and the LREE to monazite ($LREE_{Mnz}$) and multiplied by the fraction of anatectic melt at P – T retrieved from THERMOCALC to arrive at the fraction of apatite and monazite dissolution (Ap_{diss} and Mnz_{diss}) required to saturate the anatectic melt in these elements (e.g. Kelsey et al., 2008):

$$Ap_{diss} = \frac{P_{Sat}}{P_{Ap}} \times melt \text{ fraction} \quad (11)$$

$$Mnz_{diss} = \frac{LREE_{Sat}}{LREE_{Mnz}} \times melt \text{ fraction} \quad (12)$$

Finally, this last calculation is repeated considering the amount of LREE contributed by the breakdown of apatite and the amount of phosphorus contributed by the breakdown of monazite:

$$Ap_{diss}^* \% = \left[\frac{P_{Sat}}{P_{Ap} + (Mnz_{diss} \times P_{Mnz})} \right] \times melt \text{ fraction} \times 100 \quad (13)$$

$$Mnz_{diss\%}^* = \left[\frac{LREE_{Sat}}{LREE_{Mnz} + (Ap_{diss} \times LREE_{Ap})} \right] \times melt\ fraction \times 100 \quad (14)$$

The result is the percentage of apatite ($Ap_{diss\%}^*$) and monazite ($Mnz_{diss\%}^*$) dissolution required to saturate the volume of anatectic melt in phosphorus and the LREE with respect to apatite and monazite. These values are plotted on a P – T grid and are contoured for closed-system scenarios.

For the open-system scenarios, the same equations are employed except that the effective bulk composition of phosphorus (P_{bulk}) and the LREE ($LREE_{Bulk}$) change along the P – T path after melt loss events. Similar to the closed system scenario, three initial bulk compositions of P_2O_5 are investigated (0.1, 0.3 and 0.5 wt.%) for open-system behaviour and the initial bulk composition of LREE is set at 150 ppm.

The equations above are used when the system contains apatite and monazite to buffer the concentration of phosphorus and the LREE in melt (P_{Sat} and $LREE_{Sat}$, respectively). In scenarios where monazite is completely consumed, the LREE concentrations of melt and apatite are determined by the partitioning expression in equation (4). In scenarios where monazite and apatite are both completely consumed, the concentrations of phosphorus and the LREE in melt are calculated from the proportion of melt in the system and the effective bulk compositions of phosphorus (P_{bulk}) and LREE ($LREE_{Bulk}$).

4. Results for a closed system

4.1 Phase relations

The P – T pseudosection for the amphibolite-facies metapelite is shown in Figure 1. The mineral assemblages are typical of those found in metasedimentary migmatites at the amphibolite and granulite facies. Ilmenite and plagioclase are stable in each suprasolidus field across the modelled P – T range. Rutile is restricted to high pressures ($P > 10$ – 11.5 kbar). The important peritectic minerals in the modelled composition are K-feldspar, garnet, and cordierite. K-feldspar is stable at $T > 660$ – 770°C with increasing pressure. Garnet is stable across the diagram except at $T < 700^\circ\text{C}$ at $P < 5.5$ kbar. Cordierite

248 is stable at $P < 4\text{--}7$ kbar with increasing temperature. Orthopyroxene is not stable in the modelled $P\text{--}T$
249 range.

250 The four important partial melting reactions for the metapelite include, from low T to high T : (1)
251 the consumption of a minimal amount of free H_2O to produce melt at the wet solidus, (2) the breakdown
252 of muscovite to produce peritectic K-feldspar, which is represented by a narrow low-variance field that
253 extends from the wet solidus at ~ 4 kbar up to 780°C at 12 kbar, (3) the progressive consumption of
254 biotite to produce peritectic garnet at higher pressure or peritectic cordierite at lower pressure, and (4)
255 the consumption of quartz and feldspar to produce progressively drier melt at temperatures above biotite
256 exhaustion ($>900^\circ\text{C}$). Quartz is predicted to be completely consumed at $820\text{--}950^\circ\text{C}$ with increasing
257 pressure. These melting reactions are similar to those documented in experimental melting studies
258 (Clemens, 2006) and in migmatites terranes (e.g. Brown, 2013) for metasedimentary rocks.

259

260 *4.2 Melt proportions and composition*

261 The proportion of melt predicted at $P\text{--}T$ in the metapelite is shown in Figure 2a for a closed
262 system. Melt isopleths are positively sloping to nearly vertical. Therefore, melting can proceed through
263 heating or decompression in a closed-system scenario. However, this may not be the case for open-
264 system scenarios (e.g. Yakymchuk and Brown, 2014a). Note that the melt contours are not evenly
265 spaced, which indicates that melt production is non-linear with heating.

266 The composition of the melt varies across the pseudosection and the important compositional
267 parameters for apatite and monazite dissolution are shown in Figure 2b–d. Melt ASI increases as
268 temperature increases (Fig. 2b), which reflects the progressive breakdown of muscovite and biotite.
269 Contours for ASI are positively sloping on the pseudosection in the stability field of an aluminosilicate
270 mineral (sillimanite and kyanite) and are negatively sloping below the sillimanite-out line in the low P
271 and high T portion of the pseudosection. This indicates that the solubility of apatite is expected to
272 increase during prograde metamorphism and anatexis for the WL94 model of apatite solubility.

273 The weight fraction of SiO_2 in the anatectic melt generally increases up temperature, with the
274 exception of low P and high T where it decreases up temperature due to the exhaustion of quartz (Fig.
275 2c). Contours for the weight fraction of SiO_2 have moderate positive slopes at low T and shallower
276 slopes at low P and high T (Fig. 2c). An increase in SiO_2 is expected to decrease the solubility of apatite
277 in melt and an increase in temperature increases the solubility of apatite (e.g. Harrison and Watson,
278 1984).

279 Contours of the weight fraction of H_2O in melt have positive slopes across the diagram and the
280 values decrease up temperature (Fig. 2d). The decrease of the H_2O content of melt during heating
281 reflects the increasing contribution of anhydrous reactants in melt-producing reactions during heating.
282 For example, after biotite is exhausted, melting proceeds mostly through the breakdown of quartz and
283 feldspar. Based on the solubility equation of Stepanov et al. (2012), monazite solubility is expected to
284 decrease in drier melt. However, this decrease in solubility may be counteracted during prograde
285 metamorphism by the increase in temperature (e.g. Yakymchuk and Brown, 2014b).

286

287 *4.3 Phosphorus and LREE saturation of the melt*

288 The saturation values of phosphorus (P_{Sat}) and the LREE (LREE_{Sat}) for the metapelite over the
289 modelled P – T range are shown in Figure 3. For apatite, the HW84 solubility equation predicts P_2O_5
290 concentrations of up to 0.18 wt.% at 950°C and the saturation concentrations are mostly temperature
291 dependent (Fig. 3a). By contrast, the WL94 model predicts substantially higher concentrations of up to
292 ~0.60 wt.% P_2O_5 at 950°C and they are strongly correlated with melt ASI (Figs 2b, 3b). Note that the
293 WL94 model does not work for ASI values < 1.1 (Fig. 2b). Therefore, this model is not applicable for the
294 high P and low T portion of the diagram (Fig. 3b). LREE saturation contours are nearly vertical, which
295 indicates that they are mostly sensitive to temperature (Fig. 3c). Monazite solubility is also sensitive to
296 the concentration of H_2O in melt, which decreases during heating for the modelled composition (Fig.
297 2d).

298

299 4.4 Apatite behaviour

300 The two different apatite solubility equations (HW84 vs. WL94; equations 1 and 2) yield
301 different results for the behaviour of apatite. For the HW84 solubility equation, apatite dissolution
302 contours are positively sloping and nearly vertical (Fig. 4b, d, f); this is similar to the slope of melt
303 contours in Fig. 2a. The amount of apatite dissolution is non-linear and the spacing of contours
304 decreases at progressively higher temperatures. At 9 kbar and 900°C, the predicted amounts of apatite
305 dissolution for the three modelled bulk compositions vary from to 45% for low P₂O₅ concentrations
306 (0.10 wt.%), 18% for intermediate concentrations (0.30 wt.% P₂O₅), and 10% at higher concentrations
307 (0.50 wt.% P₂O₅). Complete apatite dissolution is predicted only for the low P₂O₅ concentration at
308 $T > 930^{\circ}\text{C}$ at $P < 9$ kbar (Fig. 4f).

309 For the WL94 model, the dissolution contours are strongly correlated with melt ASI (Fig. 2b)
310 and are positively sloping at low T as well as at high P and negatively sloping at $P < 5$ kbar at high T
311 (Fig. 4a, c, e). The transition from positive to negative slope generally coincides with the sillimanite-out
312 field boundary (Fig. 1). When compared with the results with the HW84 solubility equation, the WL94
313 model predicts more apatite dissolution across the investigated P – T range (Fig. 4). For both apatite
314 dissolution models, apatite is predicted to grow during cooling and melt crystallization. The breakdown
315 of monazite contributes minimal phosphorus to the melt (e.g. London et al., 1999; Garcia-Arias et al.,
316 2012) and the dissolution of apatite is essentially independent of any phosphorus released from monazite
317 breakdown.

318

319 4.5 Monazite behaviour

320 The predicted stability of monazite at suprasolidus conditions is shown in Fig. 5 for three closed-
321 system scenarios. For the first scenario, LREE liberated from apatite breakdown is not considered (e.g.
322 Kelsey et al., 2008; Yakymchuk and Brown, 2014b). For the second scenario, apatite breakdown is

323 modelled using the HW84 solubility equation and LREE is partitioned between apatite and melt using
324 equation 4. The third scenario is similar to the second except that the WL94 solubility equation for
325 apatite is used.

326 All three scenarios yield positively sloping contours for monazite dissolution that are kinked in
327 the narrow low-variance muscovite–K-feldspar and sillimanite–cordierite fields (Figs 1, 5); this reflects
328 the large quantity of melt produced in these relatively narrow fields, which requires a large proportion of
329 monazite dissolution to maintain melt saturation in the LREE. The main difference between the
330 monazite-only model (scenario 1) and the models that include LREE partitioning between apatite and
331 melt (scenarios 2 and 3) is that the dissolution contours for monazite are shifted down temperature by
332 ~10–20°C (Fig. 5b, c). This reflects the reduced effective bulk composition of LREE in the system
333 available to monazite due to the sequestration of some LREE into apatite.

334

335 **5. Results for an open system**

336 The reaction sequence for the modelled isobaric heating paths at 9 kbar and 6 kbar are compared for
337 closed and open-system scenarios in Figure 6a–d. For the 9 kbar heating path, five melt loss events
338 occur during heating resulting in the extraction of 30 mol.% melt from the system. The quantity of melt
339 produced in the open-system scenario (32 mol.%) is roughly half of the amount generated in a closed
340 system (60 mol.%). For the 6 kbar heating path, six melt loss events are predicted to occur during
341 heating and the quantity of melt produced in the open-system scenario (39 mol.%) is also roughly half of
342 the amount generated in a closed system (66 mol.%). While there are subtle differences in the reaction
343 sequence and stability of minerals between closed- and open-system scenarios (e.g. the stability of
344 quartz to higher *T* in open-system scenarios), the principle effect of progressive melt loss is the reduced
345 fertility of the residuum. The differences of important melt composition variables for apatite and
346 monazite solubility in melt (ASI, SiO₂, H₂O) are negligible (<2% difference) between the closed- and

open-system models at the same P – T condition. Therefore, Figures 2b–d and 3 are broadly applicable for evaluating both closed- and open-system behaviour during isobaric heating.

For the open-system scenarios, bulk rock values of P_2O_5 and LREE change along the P – T path at each melt loss event for both modelled P – T paths (Fig. 6e–l). Changes in the concentrations of P_2O_5 and LREE in the melt phase are shown in Figure 7 for calculations using the WL94 apatite solubility equation. If concentrations of P_2O_5 and LREE in melt (Fig. 7)—which are equivalent to the saturation concentrations of phosphorus (P_{Sat}) and the LREE ($LREE_{Sat}$) if apatite and monazite are present—are higher than the effective compositions (P_{bulk} and $LREE_{Bulk}$), then the residue will become depleted in these elements after melt extraction (Fig. 6e–l). Extraction of melt with lower concentrations of phosphorus and the LREE will result in elevated residuum concentrations of these elements. For example, the P_2O_5 value of the residue increases after melt loss events for the isobaric heating path at 9 kbar and the starting compositions of 0.3 and 0.5 wt.% P_2O_5 (Fig. 6e, f) because the melt contains lower concentrations of these elements (Fig. 7a). By contrast, there is a step-wise decrease in the P_2O_5 value of the residuum for the isobaric heating path at 6 kbar and starting P_2O_5 concentration of 0.1 wt.% using the WL94 solubility equation (Fig. 6g) because the P_2O_5 concentration of the extracted melt is >0.1 wt.% (Fig. 7c). The concentration of LREE in melt (Fig. 7b, d) is higher than the bulk composition at each melt loss event and the residue becomes more depleted in LREE after each melt loss event (Fig. 6i–l).

The amounts of apatite and monazite dissolution along the isobaric heating paths are compared for closed- and open-system scenarios in Figure 8. The modes of apatite and monazite decrease during heating for both scenarios. Melt loss increases the stability of apatite and monazite to higher temperatures in open systems relative to the closed-system scenarios.

The stability of apatite to higher temperatures for the open-system scenarios reflects both the progressive enrichment of the residue in P_2O_5 in most cases (Fig. 6e–g) and the reduced fertility of the metapelite during progressive melt extraction along the isobaric heating path (Fig. 6a–d). An extreme

case is for a starting P_2O_5 value of 0.1 wt.% using the WL94 solubility equation for the isobaric heating path at 9 kbar. For the closed system, apatite is expected to be completely consumed at $\sim 880^\circ C$ (Fig. 8a). For the open system, less than 30% of the amount of apatite present at the solidus is predicted to be consumed by $950^\circ C$ (Fig. 8a).

The increased stability of monazite to higher temperatures mainly reflects the reduced fertility of the metapelite. The decrease in the LREE value of the residue during melt loss only has a subordinate effect. The difference between closed- and open-system scenarios for monazite stability are apparent at higher temperatures where there is significant divergence between the dissolution curves for the different scenarios (Fig. 8e–h). Again, this reflects the reduced fertility of the metapelite after several melt-loss events (Fig. 6a–d).

6. Discussion

6.1. Limitations of modelling

The limitations of the modelling approach applied here for monazite behaviour are detailed in Kelsey et al. (2008), Yakymchuk and Brown (2014), and are also discussed by Taylor et al. (2016). There are some additional limitations that apply to modelling apatite behaviour in this study. No adjustment was made to the average metapelite composition to account for Ca in apatite in the phase equilibria modelling. The modelling does not consider fluorine and chlorine, which can be important for apatite stability in some magmatic, hydrothermal and metamorphic systems (e.g. Bingen et al. 1996; Piccoli and Candela, 2002; Webster and Piccoli, 2015). Model apatite is assumed to have homogeneous zoning in trace elements, although natural apatite can be zoned (e.g. Rakovan and Reeder, 1996; Smith and Yardley, 1999; Yang and Rivers, 2002).

Apatite and monazite are modelled here as stoichiometric minerals (Apatite: $Ca_5(PO_4)_3(OH)$, and monazite $[LREE]PO_4$). Consequently, the value of X_{mnz}^{LREE} in the Stepanov et al. (2012) monazite solubility equation (equation 3 above) is 1. This approach ignores potentially important components

397 such as Th, U and Pb. Thorium concentrations in monazite usually range between 1–10 wt.% (e.g.
398 Stepanov et al., 2012), but can be up to 30 wt.% of some monazite (e.g. Catlos, 2013). A consequence of
399 considering Th substitution in monazite for the modelling presented here is that LREE saturation value
400 of equation (3) will decrease. Consequently, less LREE will be allocated to apatite in the bulk
401 composition ($LREE_{Ap}$), which is controlled by the LREE concentration of the melt (equation 4). The net
402 effect is that high-Th monazite is expected to persist to higher temperatures than Th-poor monazite.

403 Various values of X_{mnz}^{LREE} assuming Th substitution in monazite are investigated along the open-
404 system isobaric heating paths using the WL94 apatite solubility equation in Fig. 9. Note that the
405 calculations for monazite dissolution require an adjustment of the LREE composition of monazite
406 (Mnz_{LREE}) to account for Th substitution. For the 6 kbar isobaric heating path, the stability of monazite is
407 increased by ~50°C at high temperature for an X_{mnz}^{LREE} value of 0.7 (Fig. 9b). The high-temperature limit
408 of monazite stability is more significant for the 9 kbar heating path where ~30% of the monazite
409 originally present at the solidus is preserved at 950°C for an X_{mnz}^{LREE} value of 0.7 (Fig. 9a).

410 Although apatite and monazite are the main mineral repositories of phosphorus in most suprasolidus
411 metamorphic rocks, garnet and plagioclase can accommodate significant quantities of phosphorus (e.g.
412 Pyle and Spear, 2002; Villaseca et al., 2003; Kohn and Malloy, 2004; Dumond et al., 2015). For
413 example, ultrahigh-temperature granulites from Connecticut contain garnet with P_2O_5 concentrations up
414 to 0.4 wt.% (Ague and Eckert, 2012; Axler and Ague, 2015a, b). In the modelling here, the result of
415 garnet accommodating phosphorus is a reduction of phosphorus in the system allocated to apatite and an
416 increase in the amount of apatite dissolution required to maintain phosphorus saturation of melt. For the
417 modelled prograde heating paths and for both closed- and open-system scenarios (Fig. 6a–d), the
418 increasing mode of garnet would result in the partitioning of a greater proportion of the bulk phosphorus
419 into garnet as temperature increased. This would decrease the amount of phosphorus allocated to apatite
420 and may accelerate the consumption of apatite. However, the same heating paths would also consume

plagioclase, which may release some phosphorus into the system and contribute to phosphorus saturation of the melt.

The consequences of garnet and plagioclase behaviour on the modes of apatite (and monazite) may need to be considered in some cases where these major minerals are important hosts for phosphorus. However, these minerals are generally expected to only contain a few percent of the total phosphorus budget for metamorphic rocks. For example, a simple mass balance calculation by Spear and Pyle (2002) estimate that a rock with 10 modal % garnet with 200 ppm phosphorus accounts for ~6% of the phosphorus budget of the entire rock. Furthermore, the covariation between LREE and phosphorus concentrations in some migmatites and anatectic granites suggests that the accessory minerals are the main control on these elements during anatexis (e.g. Watt and Harley, 1993; Zeng et al., 2005a; Brown et al., 2016).

6.2 Kinetic controls on apatite and monazite dissolution

The modelling presented here assumes equilibration of the anatectic melt in major and trace elements. However, kinetics may inhibit apatite and monazite dissolution during partial melting, which may lead to undersaturation of the melt in phosphorus and the LREE (e.g. Bea et al., 1994; Ayers and Harris, 1997; Zeng et al., 2005a). The kinetics of apatite and monazite dissolution into anatectic melt are sensitive to grain size, temperature and the H₂O content of the melt (e.g. Rapp and Watson, 1986; Harrison and Watson, 1984). For a melt with 6 wt.% H₂O, it is predicted to take a maximum of ~1 ky at 750°C and <100 years at 950°C to dissolve a 500µm diameter apatite crystal (Fig. 10a; Harrison and Watson, 1984). For a dry melt, a 500µm diameter apatite crystal is expected to take >10 My to completely dissolve at 950°C (Fig. 10a). However, a completely dry melt composition is unrealistic for the modelled metapelite composition for both closed- and open-system scenarios; the lowest modelled H₂O content of melt in both scenarios is ~2 wt.% H₂O.

445 For monazite, Rapp and Watson (1986) evaluated the maximum time it takes to completely
446 consume monazite grains of various sizes. It is expected to take ~1 My at 750°C and <100 years at
447 950°C to completely dissolve a 50µm grain of monazite in a melt with 6 wt.% H₂O (Fig. 10b). For a
448 drier melt with 1 wt.% H₂O, the same monazite grain is predicted to take >1 Gy at 750°C and ~1 My at
449 950°C to completely dissolve (Fig. 10b).

450 For the modelled metapelite composition, the H₂O content of the melt decreases up temperature
451 for closed-system and open-system scenarios. The difference in the predicted H₂O content of melt
452 between these different scenarios is negligible (<0.2 wt.%) at the same *P–T* condition. Although drier
453 melt compositions are expected to hinder monazite dissolution, this is counteracted by the increase in
454 temperature. For example, the H₂O content of melt at 9 kbar and 750°C is 9.5 wt.% (Fig. 2d). At these
455 *P–T* conditions a 50µm grain of monazite is predicted to take <1 My to completely dissolve (Fig. 10b).
456 A similar duration is expected for a melt with ~3 wt.% H₂O at 9 kbar and 950°C (Fig. 10b).

457 The timescales of most regional high-temperature metamorphic processes range from millions to
458 tens of millions of years (e.g. Hermann and Rubatto, 2003; Clark et al., 2011; Harley, 2016). In general,
459 the timescales for apatite and monazite dissolution for the modelled average metapelite are expected to
460 be <1 My for anatectic melt with realistic H₂O concentrations (>1 wt.%). Therefore, kinetics are not
461 expected to hinder apatite and monazite dissolution in long-lived migmatite terranes except possibly for
462 large grains of apatite (> 500 µm) and monazite (> 50 µm) or if non-equilibrated melt is periodically
463 extracted at <1 My intervals (e.g. Sawyer, 1991; Ayers and Harris, 1997).

464

465 6.3 Prograde monazite growth in migmatites

466 Theoretical modelling of monazite behaviour in suprasolidus metamorphic rocks predicts that
467 prograde monazite growth above the solidus is unlikely because monazite breakdown is required for
468 LREE-saturation of the increasing amount of anatectic melt generated during heating (e.g. Kelsey et al.,
469 2008; Spear and Pyle, 2010; Yakymchuk and Brown, 2014b). Similarly, the modelling here suggests

470 that apatite will be consumed during heating and partial melting and grow as melt crystallizes during
471 cooling to the solidus for both closed- and open-system scenarios. Because apatite is generally stable to
472 higher temperatures than monazite (Fig. 8) and partitions some of the LREE in the system, the result is a
473 reduced stability of monazite when compared with an apatite-free model of monazite dissolution.
474 Therefore, no new monazite is predicted to grow during heating and partial melting including from the
475 breakdown of LREE-rich apatite in an equilibrated system.

476 If monazite is exhausted during prograde metamorphism, the concentration of LREE in the melt
477 will be governed by the partitioning of LREE between it and apatite and the LREE concentration of the
478 melt will decrease with further partial melting due to dilution (e.g. Klimm et al., 2008). The melt will
479 remain undersaturated with respect to monazite for the remainder of the prograde path and no new
480 monazite growth is expected.

481 Johnson et al. (2015) suggested that the breakdown of LREE-rich apatite during prograde
482 metamorphism may have liberated enough LREE to oversaturate the melt and grow monazite due to an
483 increase in melt ASI during progressive partial melting. However, the results of the modelling here
484 suggest that the increase in ASI appears to be counteracted by the non-linear increase in the generation
485 of melt during heating in a closed system (Fig. 6a, c), which requires progressively more apatite (and
486 monazite) dissolution to achieve phosphorus saturation of the melt. For an open system, the fertility of
487 the system decreases after melt loss (Fig. 6b, d), but monazite is still expected to be consumed during
488 heating to saturate the melt in LREE and no new growth is expected (Fig. 8e–h). Both of the modelled
489 scenarios assume that the melt is homogeneous and saturated in phosphorus (with respect to apatite)
490 during the entire metamorphic evolution except at P – T conditions above apatite exhaustion.

491 One of the key assumptions of a homogeneous melt composition in this study may not be valid if
492 there is sluggish diffusion of phosphorus and/or LREE away from apatite during dissolution. Wolf and
493 London (1995) suggested that monazite can grow along the apatite–melt interface due to local
494 oversaturation of LREE during apatite breakdown. This is because the diffusivities of REE and

495 phosphorus are generally similar (e.g. Rapp and Watson, 1986) whereas the diffusivity of calcium is
496 approximately two orders of magnitude higher (Harrison and Watson, 1984). Consequently, a narrow
497 region of melt adjacent to apatite may become enriched in LREE and phosphorus relative to the
498 surrounding melt and may be locally saturated with respect to monazite. This has been called ‘pileup’
499 where diffusion away from the crystal interface is too slow to keep up with melt equilibration (e.g.
500 Green and Watson, 1982; Harrison and Watson, 1984). Therefore, although monazite crystallization
501 from apatite breakdown is not predicted by the phase equilibria modelling in this study for both closed-
502 and open-system scenarios, new monazite growth may occur in systems where the melt composition is
503 not homogeneous.

504 The extent of melting may have important implications for melt homogenization. Partial melting
505 begins at reactant grain junctions (e.g. Sawyer, 1999; Acosta-Vigil et al., 2006) in migmatites and melt
506 initially accumulates in isolated pockets. Further heating will increase the proportion of melt in the
507 system and the migmatite will eventually develop an interconnected melt network that physically links
508 individual melt pockets (e.g. Holness et al., 2011; Sawyer, 2014). This may lead to melt extraction from
509 the system as modelled above for the open-system scenarios. Isolated melt pockets have the potential to
510 become oversaturated with respect to monazite if LREE-rich apatite is in contact with the melt and
511 diffusion is sluggish. However, once an interconnected melt network develops, this will drive the system
512 towards homogenization and the preservation of oversaturated melt with respect to monazite (and
513 apatite) becomes less likely. Rosenberg and Handy (2005) estimated that melt interconnectivity may
514 occur when the melt proportion reaches ~7 vol.% in a static system. This value may be much less in a
515 dynamic system undergoing syn-anatectic deformation (Vigneresse and Burg, 2000; Brown, 2013),
516 which is more applicable to regional migmatite terranes that are generally very weak during anatexis
517 (e.g. Diener et al., 2014). For a closed system, the preservation of prograde monazite from the
518 breakdown of apatite is more likely in migmatites that have not undergone extensive partial melting or
519 where melt does not form an interconnected network. For open-system scenarios, isolated melt pockets

520 may become temporarily connected, but in general they evolve independent from each other and have a
521 greater likelihood of reaching local LREE oversaturation with respect to monazite.

522 Monazite growth at the margin of consumed apatite has been documented in experiments (Wolf
523 and London, 1995; García-Arias et al., 2012) and in studies of metamorphic rocks (e.g. Harlov et al.,
524 2007; Finger and Krenn, 2007; Rocha et al., 2016). Wolf and London (1995) described small (~30 µm)
525 monazite grains at the apatite–melt interface in their experiments. Similarly, Harlov et al. (2007)
526 documented small (<20 µm) monazite crystals at the margin of apatite in granulites from the Variscan
527 Schwarzwald of southern Germany. García-Arias et al. (2012) found <10 µm monazite grains around
528 apatite generated during experimental melting of an Iberian orthogneiss. Considering the kinetic
529 modelling of previous studies (Fig. 10; Harrison and Watson, 1984; Wolf and London, 1995), the
530 predicted dissolution times for these grain sizes are <1 My except for unrealistically dry melt
531 compositions. Therefore, prograde monazite produced through apatite breakdown is expected to be
532 consumed during melt equilibration and/or further partial melting. However, if these crystals are
533 included into a growing peritectic mineral (e.g. garnet), then they have a higher preservation potential.
534 As suggested by Wolf and London (1995), small monazite inclusions encased in minerals may reflect
535 earlier (e.g. prograde) growth whereas larger monazite grains in the matrix may grow from crystallizing
536 residual anatectic melt during cooling.

537 The reaction sequence experienced by migmatites may be an important factor in the preservation
538 of prograde suprasolidus monazite in peritectic minerals. For the modelled isobaric heating paths in
539 Figure 6, muscovite breakdown produces peritectic K-feldspar at ~700–750°C and the mode of K-
540 feldspar is expected to increase during biotite breakdown until ~810–850°C when biotite becomes
541 exhausted. For the 6 kbar isobaric heating path, cordierite grows at the expense of biotite. At
542 temperatures above biotite exhaustion, K-feldspar and cordierite are consumed. Therefore, any monazite
543 that may be generated at the margins of dissolving apatite over the interval 750–850°C has the potential
544 to be preserved in growing K-feldspar and/or cordierite, but at higher temperatures the rims of these

minerals may be resorbed, which may release some of the included monazite back into the melt. The amount of apatite dissolution over this temperature range for a closed system is small and is predicted to vary between 10% and 50% using the WL94 solubility equation for the modelled bulk P_2O_5 concentrations (Fig. 8). The amount is expected to be even less for the HW84 solubility equation and much less for the modelled open-system scenarios (Fig. 8).

Garnet is an important peritectic mineral that is commonly targeted for *in situ* monazite U–Pb geochronology (e.g. Korhonen et al., 2012; Yakymchuk et al., 2015). Similar to K-feldspar and cordierite, garnet growth is predicted to occur at the expense of biotite and is consumed during heating after biotite exhaustion in both closed- and open-system scenarios (Fig. 6). Therefore, garnet also has the potential to capture monazite derived from apatite breakdown at $T < 810\text{--}850^\circ\text{C}$ in the modelled metapelite along the 6 kbar and 9 kbar isobaric heating paths (Fig. 6). Dumond et al. (2015) suggested that some of their monazite inclusions in garnet are related to the breakdown of apatite (that now form rounded inclusions in garnet) during heating and partial melting.

One mechanism that has been proposed to produce prograde suprasolidus monazite is Ostwald ripening (e.g. Nemchin and Bodorkos, 2000). Ostwald ripening reduces the total surface free energy of a system by preferentially dissolving smaller solids and precipitating this material on existing larger solids (e.g. Tikare and Cawley, 1998). This is a possible mechanism for prograde monazite growth in suprasolidus metamorphic rocks if there is no net decrease in the mode of monazite (e.g. Yakymchuk et al., in press). However, monazite modes are expected to decrease during heating in both closed and open systems to saturate the melt in the LREE. Therefore, for the scenarios modelled here, Ostwald ripening may not be an appropriate mechanism for prograde monazite growth in an equilibrated system.

6.4 Apatite dissolution and granite chemistry

Apatite behaviour during partial melting has important implications for studies of the elemental and isotopic variability of granites (e.g. Ayers and Harris, 1997; Zeng et al., 2005a, 2005b; Farina and

570 Stevens, 2011; Farina et al., 2014). Because apatite has a higher Sm/Nd ratio than monazite and the bulk
571 rock (e.g. Bea et al., 1994; Ayers and Harris, 1997), apatite will develop a relatively more radiogenic Nd
572 isotope value than the source. Detrital apatite also maintains its Nd isotope ratio during metamorphism
573 (e.g. Hammerli et al., 2014). Therefore, the dissolution of significant amounts of apatite (e.g. non-modal
574 melting) may result in a melt with a higher Nd isotope composition compared with the source (e.g. Zeng
575 et al., 2005a, 2005b).

576 The results of the modelling presented here indicate that the proportion of apatite dissolution is
577 sensitive to the bulk P_2O_5 concentration. For high P_2O_5 bulk concentrations, apatite progressively breaks
578 down towards higher T , but some apatite is expected to survive up to UHT conditions for both closed-
579 and open-system scenarios. Consequently, melt may develop a progressively more radiogenic Nd
580 isotope signature as melting proceeds. By contrast, if apatite is exhausted in a low P_2O_5 bulk
581 composition, the melt will become progressively less radiogenic due to the breakdown of minerals with
582 lower Sm/Nd ratios during further partial melting. However, monazite is expected to be the main
583 repository for Nd in pelites (e.g. Hammerli et al., 2014) and its breakdown will likely exert a first-order
584 control on the Nd isotope ratio of the melt.

585 The equilibrium modelling here predicts that melt remains saturated with phosphorus until
586 apatite is exhausted at high T . After the complete exhaustion of apatite during partial melting, additional
587 heating is predicted to generate melt that is undersaturated with respect to phosphorus. In addition to
588 apatite exhaustion, there are two important mechanisms that can produce undersaturated anatectic melt,
589 including: (1) inhibited accessory mineral dissolution due to kinetic factors (e.g. Harrison and Watson,
590 1984; Rapp and Watson, 1986) and (2) non-Henrian behaviour of trace elements during melt–solid
591 partitioning (Bea, 1996). Together, these mechanisms can result in a melt that is undersaturated in
592 phosphorus with respect to apatite, which may hinder prograde monazite growth above the solidus.

593 An additional factor that impacts apatite dissolution is the sequestration of grains included in
594 other minerals away from the reaction volume (e.g. Watson et al., 1989; Watt and Harley, 1993). In this

case, the effective bulk composition will be reduced and the high-temperature stability of the accessory mineral will be lowered (e.g. Yakymchuk and Brown, 2014b). This has the potential to produce undersaturated melt at temperatures below those predicted by using a measured P_2O_5 bulk concentration and reduce the stability of apatite in the reacting volume of the rock (e.g. along grain boundaries). The exhaustion of apatite also limits the use of apatite saturation temperatures to rocks where apatite was still present at high temperature.

7. Conclusions

Apatite is predicted to progressively dissolve into anatectic melt during prograde suprasolidus metamorphism in metapelites for both closed- and open-system scenarios. The stability of apatite is primarily a function of melt composition, temperature and the bulk rock concentration of P_2O_5 and whether melt loss is considered. For low bulk rock P_2O_5 concentrations, apatite may be completely dissolved at UHT conditions whereas some apatite may survive at these conditions for high P_2O_5 bulk rock concentrations. Melt loss generally extends the stability of apatite and monazite to higher temperature. Partitioning of LREE into apatite reduces the high-temperature stability of monazite and only minor prograde monazite growth may occur at the apatite–melt interface due to local oversaturation of the melt in LREE. This monazite may be preserved as inclusions in peritectic minerals that shielded them from further dissolution, but prograde suprasolidus monazite is less likely to survive in the matrix of the rock.

Acknowledgements

This work was partially funded by a Discovery Grant from the National Sciences and Engineering Research Council of Canada. I thank DE Kelsey and an anonymous reviewer for detailed and constructive reviews and M Scambelluri for his editorial work.

References

- Acosta-Vigil, A., London, D., Morgan, G.B., 2006. Experiments on the kinetics of partial melting of a leucogranite at 200 MPa H₂O and 690–800 C: compositional variability of melts during the onset of H₂O-saturated crustal anatexis. *Contributions to Mineralogy and Petrology* 151, 539–557.
- Ague, J.J., 1991. Evidence for major mass transfer and volume strain during regional metamorphism of pelites. *Geology* 19, 855–858.
- Ague, J.J., Eckert, J.O., 2012. Precipitation of rutile and ilmenite needles in garnet: Implications for extreme metamorphic conditions in the Acadian Orogen, U.S.A. *American Mineralogist* 97, 840–855.
- Axler, J.A., Ague, J.J., 2015a. Oriented multiphase needles in garnet from ultrahigh-temperature granulites, Connecticut, U.S.A. *American Mineralogist* 100, 2254–2271.
- Axler, J.A., Ague, J.J., 2015b. Exsolution of rutile or apatite precipitates surrounding ruptured inclusions in garnet from UHT and UHP rocks. *Journal of Metamorphic Geology* 33, 829–848.
- Ayres, M., Harris, N., 1997. REE fractionation and Nd-isotope disequilibrium during crustal anatexis: constraints from Himalayan leucogranites. *Chemical Geology* 139, 249–269.
- Bea, F., 1996a. Controls on the trace element composition of crustal melts. *Geological Society of America Special Papers* 315, 33–41.
- Bea, F., 1996b. Residence of REE, Y, Th and U in granites and crustal protoliths; implications for the chemistry of crustal melts. *Journal of Petrology* 37, 521–552.
- Bea, F., Fershtater, G., Corretgé, L.G., 1992. The geochemistry of phosphorus in granite rocks and the effect of aluminium. *Lithos* 29, 43–56.
- Bea, F., Pereira, M.D., Stroh, A., 1994. Mineral/leucosome trace-element partitioning in a peraluminous migmatite (a laser ablation-ICP-MS study). *Chemical Geology* 117, 291–312.
- Bingen, B., Demaiffe, D., Hertogen, J., 1996. Redistribution of rare earth elements, thorium, and uranium over accessory minerals in the course of amphibolite to granulite facies metamorphism: The role of apatite and monazite in orthogneisses from southwestern Norway. *Geochimica et Cosmochimica Acta* 60, 1341–1354.
- Blereau, E., Clark, C., Taylor, R.J.M., Johnson, T.E., Fitzsimons, I.C.W., Santosh, M., 2016. Constraints on the timing and conditions of high-grade metamorphism, charnockite formation and fluid-rock interaction in the Trivandrum Block, southern India. *Journal of Metamorphic Geology* 34, 527–549.
- Brown, M., 2013. Granite: From genesis to emplacement. *Geological Society of America Bulletin* 125, 1079–1113.
- Brown, C.R., Yakymchuk, C., Brown, M., Fanning, C.M., Korhonen, F.J., Piccoli, P.M., Siddoway, C.S., 2016. From Source to Sink: Petrogenesis of Cretaceous Anatectic Granites from the Fosdick Migmatite–Granite Complex, West Antarctica. *Journal of Petrology* 57, 1241–1278.
- Catlos, E.J., 2013. Review: Versatile Monazite: resolving geological records and solving challenges in materials science. Generalizations about monazite: Implications for geochronologic studies. *American Mineralogist* 98, 819–832.
- Clark, C., Fitzsimons, I.C.W., Healy, D., Harley, S.L., 2011. How does the continental crust get really hot? *Elements* 7, 235–240.
- Clemens, J.D., 2006. Melting of the continental crust: fluid regimes, melting reactions, and source-rock fertility. In: Brown, M. and Rushmer, T. (eds), *Evolution and Differentiation of the Continental Crust*. Cambridge University Press.
- Diener, J.F.A., Fagereng, Å., 2014. The influence of melting and melt drainage on crustal rheology during orogenesis. *Journal of Geophysical Research: Solid Earth* 119, 6193–6210.
- Duc-Tin, Q., Keppler, H., 2015. Monazite and xenotime solubility in granitic melts and the origin of the lanthanide tetrad effect. *Contributions to Mineralogy and Petrology* 169, 1–26.

- 669 Dumond, G., Goncalves, P., Williams, M.L., Jercinovic, M.J., 2015. Monazite as a monitor of melting,
670 garnet growth and feldspar recrystallization in continental lower crust. *Journal of Metamorphic*
671 *Geology* 33, 735-762.
- 672 Farina, F., Dini, A., Rocchi, S., Stevens, G., 2014. Extreme mineral-scale Sr isotope heterogeneity in
673 granites by disequilibrium melting of the crust. *Earth and Planetary Science Letters* 399, 103-
674 115.
- 675 Farina, F., Stevens, G., 2011. Source controlled $^{87}\text{Sr}/^{86}\text{Sr}$ isotope variability in granitic magmas: the
676 inevitable consequence of mineral-scale isotopic disequilibrium in the protolith. *Lithos* 122, 189-
677 200.
- 678 Farley, K.A., Stockli, D.F., 2002. (U-Th)/He dating of phosphates: Apatite, monazite, and xenotime.
679 *Reviews in Mineralogy and Geochemistry* 48, 559-577.
- 680 Finger, F., Krenn, E., 2007. Three metamorphic monazite generations in a high-pressure rock from the
681 Bohemian Massif and the potentially important role of apatite in stimulating polyphase monazite
682 growth along a PT loop. *Lithos* 95, 103-115.
- 683 García-Arias, M., Corretgé, L.G., Castro, A., 2012. Trace element behavior during partial melting of
684 Iberian orthogneisses: An experimental study. *Chemical Geology* 292, 1-17.
- 685 Green, T.H., Watson, E.B., 1982. Crystallization of apatite in natural magmas under high pressure,
686 hydrous conditions, with particular reference to 'Orogenic' rock series. *Contributions to*
687 *Mineralogy and Petrology* 79, 96-105.
- 688 Guevara, V.E., Caddick, M.J., 2016. Shooting at a moving target: phase equilibria modelling of high-
689 temperature metamorphism. *Journal of Metamorphic Geology* 34, 209-235.
- 690 Hacker, B.R., Kylander-Clark, A.R.C., Holder, R., Andersen, T.B., Peterman, E.M., Walsh, E.O.,
691 Munnikhuis, J.K., 2015. Monazite response to ultrahigh-pressure subduction from U-Pb dating
692 by laser ablation split stream. *Chemical Geology* 409, 28-41.
- 693 Hammerli, J., Kemp, A.I.S., Spandler, C., 2014. Neodymium isotope equilibration during crustal
694 metamorphism revealed by in situ microanalysis of REE-rich accessory minerals. *Earth and*
695 *Planetary Science Letters* 392, 133-142.
- 696 Harley, S.L., 2016. A matter of time: The importance of the duration of UHT metamorphism. *Journal of*
697 *Mineralogical and Petrological Sciences* 111, 50-72.
- 698 Harlov, D.E., Marschall, H.R., Hanel, M., 2007. Fluorapatite-monazite relationships in granulite-facies
699 metapelites, Schwarzwald, southwest Germany. *Mineralogical Magazine* 71, 223-234.
- 700 Harrison, T.M., Watson, E.B., 1984. The behavior of apatite during crustal anatexis: equilibrium and
701 kinetic considerations. *Geochimica et Cosmochimica Acta* 48, 1467-1477.
- 702 Hermann, J., Rubatto, D., 2003. Relating zircon and monazite domains to garnet growth zones: age and
703 duration of granulite facies metamorphism in the Val Malenco lower crust. *Journal of*
704 *Metamorphic Geology* 21, 833-852.
- 705 Holland, T.J.B., Powell, R., 2011. An improved and extended internally consistent thermodynamic
706 dataset for phases of petrological interest, involving a new equation of state for solids. *Journal of*
707 *Metamorphic Geology* 29, 333-383.
- 708 Holness, M.B., Cesare, B., Sawyer, E.W., 2011. Melted rocks under the microscope: microstructures
709 and their interpretation. *Elements* 7, 247-252.
- 710 Johnson, T.E., Clark, C., Taylor, R.J.M., Santosh, M., Collins, A.S., 2015. Prograde and retrograde
711 growth of monazite in migmatites: An example from the Nagercoil Block, southern India.
712 *Geoscience Frontiers* 6, 373-387.
- 713 Kelsey, D.E., Clark, C., Hand, M., 2008. Thermobarometric modelling of zircon and monazite growth in
714 melt - bearing systems: Examples using model metapelitic and metapsammitic granulites.
715 *Journal of Metamorphic Geology* 26, 199-212.

716 Kelsey, D.E., Hand, M., 2015. On ultrahigh temperature crustal metamorphism: phase equilibria, trace
 717 element thermometry, bulk composition, heat sources, timescales and tectonic settings.
 718 *Geoscience Frontiers* 6, 311-356.
 719 Klimm, K., Blundy, J.D., Green, T.H., 2008. Trace Element Partitioning and Accessory Phase
 720 Saturation during H₂O-Saturated Melting of Basalt with Implications for Subduction Zone
 721 Chemical Fluxes. *Journal of Petrology* 49, 523-553.
 722 Kohn, M.J., Malloy, M.A., 2004. Formation of monazite via prograde metamorphic reactions among
 723 common silicates: implications for age determinations. *Geochimica et Cosmochimica Acta* 68,
 724 101-113.
 725 Korhonen, F.J., Brown, M., Grove, M., Siddoway, C.S., Baxter, E.F., Inglis, J.D., 2012. Separating
 726 metamorphic events in the Fosdick migmatite–granite complex, West Antarctica. *Journal of*
 727 *Metamorphic Geology* 30, 165-192.
 728 Lederer, G.W., Cottle, J.M., Jessup, M.J., Langille, J.M., Ahmad, T., 2013. Timescales of partial melting
 729 in the Himalayan middle crust: insight from the Leo Pargil dome, northwest India. *Contributions*
 730 *to Mineralogy and Petrology* 166, 1415-1441.
 731 London, D., Wolf, M.B., Morgan, G.B., Garrido, M.G., 1999. Experimental silicate–phosphate
 732 equilibria in peraluminous granitic magmas, with a case study of the Albuquerque batholith at
 733 Tres Arroyos, Badajoz, Spain. *Journal of Petrology* 40, 215-240.
 734 Mayne, M.J., Moyen, J.F., Stevens, G., Kaislaniemi, L., 2016. Rcrust: a tool for calculating path-
 735 dependent open system processes and application to melt loss. *Journal of Metamorphic Geology*
 736 34, 663-682.
 737 Montel, J.-M., 1986. Experimental determination of the solubility of Ce-monazite in SiO₂-Al₂O₃-K₂O-
 738 Na₂O melts at 800 C, 2 kbar, under H₂O-saturated conditions. *Geology* 14, 659-662.
 739 Montel, J.-M., 1993. A model for monazite/melt equilibrium and application to the generation of
 740 granitic magmas. *Chemical Geology* 110, 127-146.
 741 Nemchin, A.A., Bodorkos, S., 2000. Zr and LREE concentrations in anatectic melt as a function of
 742 crystal size distributions of zircon and monazite in the source region. *Geological Society of*
 743 *America, Abstracts and Programs*, Abstract 52286.
 744 Parrish, R.R., 1990. U-Pb dating of monazite and its application to geological problems. *Canadian*
 745 *Journal of Earth Sciences* 27, 1431-1450.
 746 Piccoli, P.M., Candela, P.A., 2002. Apatite in igneous systems. *Reviews in mineralogy and*
 747 *geochemistry* 48, 255-292.
 748 Pichavant, M., Montel, J.-M., Richard, L.R., 1992. Apatite solubility in peraluminous liquids:
 749 Experimental data and an extension of the Harrison-Watson model. *Geochimica et*
 750 *Cosmochimica Acta* 56, 3855-3861.
 751 Pyle, J.M., Spear, F.S., Wark, D.A., 2002. Electron Microprobe Analysis of REE in Apatite, Monazite
 752 and Xenotime: Protocols and Pitfalls. *Reviews in mineralogy and geochemistry* 48, 337-362.
 753 Powell, R., Holland, T.J.B., 1988. An internally consistent dataset with uncertainties and correlations: 3.
 754 Applications to geobarometry, worked examples and a computer program. *Journal of*
 755 *Metamorphic Geology* 6, 173-204.
 756 Prowatke, S., Klemme, S., 2006. Trace element partitioning between apatite and silicate melts.
 757 *Geochimica et Cosmochimica Acta* 70, 4513-4527.
 758 Rakovan, J., Reeder, R.J., 1996. Intracrystalline rare earth element distributions in apatite: Surface
 759 structural influences on incorporation during growth. *Geochimica et Cosmochimica Acta* 60,
 760 4435-4445.
 761 Rapp, R.P., Ryerson, F.J., Miller, C.F., 1987. Experimental evidence bearing on the stability of monazite
 762 during crustal anatexis. *Geophysical Research Letters* 14, 307-310.

763 Rapp, R.P., Watson, E.B., 1986. Monazite solubility and dissolution kinetics: implications for the
764 thorium and light rare earth chemistry of felsic magmas. *Contributions to Mineralogy and*
765 *Petrology* 94, 304-316.

766 Rocha, B.C., Moraes, R., Möller, A., Cioffi, C.R., Jercinovic, M.J., 2016. Timing of anatexis and melt
767 crystallization in the Socorro–Guaxupé Nappe, SE Brazil: Insights from trace element
768 composition of zircon, monazite and garnet coupled to U Pb geochronology. *Lithos*. Doi:
769 10.1016/j.lithos.2016.05.020.

770 Rosenberg, C.L., Handy, M.R., 2005. Experimental deformation of partially melted granite revisited:
771 implications for the continental crust. *Journal of Metamorphic Geology* 23, 19-28.

772 Sawyer, E.W., 1991. Disequilibrium melting and the rate of melt–residuum separation during
773 migmatization of mafic rocks from the Grenville Front, Quebec. *Journal of Petrology* 32, 701-
774 738.

775 Sawyer, E.W., 1999. Criteria for the recognition of partial melting. *Physics and Chemistry of the Earth,*
776 *Part A: Solid Earth and Geodesy* 24, 269-279.

777 Sawyer, E.W., 2014. The inception and growth of leucosomes: microstructure at the start of melt
778 segregation in migmatites. *Journal of Metamorphic Geology* 32, 695-712.

779 Smith, M.P., Yardley, B.W.D. 1999. Fluid evolution during metamorphism of the Otago Schist, New
780 Zealand: (II) Influence of detrital apatite on fluid salinity. *Journal of Metamorphic Geology* 17,
781 187-193.

782 Spear, F.S., Pyle, J.M., 2002. Apatite, monazite, and xenotime in metamorphic rocks. *Reviews in*
783 *mineralogy and geochemistry* 48, 293-335.

784 Spear, F.S., Pyle, J.M., 2010. Theoretical modeling of monazite growth in a low-Ca metapelite.
785 *Chemical Geology* 273, 111-119.

786 Spear, F.S., Kohn, M.J., Cheney, J.T., 1999. P-T paths from anatectic pelites. *Contributions to*
787 *Mineralogy and Petrology* 134, 17-32.

788 Stepanov, A.S., Hermann, J., Rubatto, D., Rapp, R.P., 2012. Experimental study of monazite/melt
789 partitioning with implications for the REE, Th and U geochemistry of crustal rocks. *Chemical*
790 *Geology* 300, 200-220.

791 Taylor, R.J.M., Kirkland, C.L., Clark, C., 2016. Accessories after the facts: Constraining the timing,
792 duration and conditions of high-temperature metamorphic processes. *Lithos*.

793 Taylor, S.R., McLennan, S.M., 1985. *The Continental Crust: Its composition and evolution*. Blackwell
794 Scientific, Oxford.

795 Tikare, V., Cawley, J.D., 1998. Application of the Potts model to simulation of Ostwald ripening.
796 *Journal of the American Ceramic Society* 81, 485-491.

797 Toplis, M.J., Dingwell, D.B., 1996. The variable influence of P₂O₅ on the viscosity of melts of differing
798 alkali/aluminium ratio: Implications for the structural role of phosphorus in silicate melts.
799 *Geochimica et Cosmochimica Acta* 60, 4107-4121.

800 Vigneresse, J.L., Burg, J.P., 2000. Continuous vs. discontinuous melt segregation in migmatites: insights
801 from a cellular automaton model. *Terra Nova* 12, 188-192.

802 Villaseca, C., Martín Romera, C., De la Rosa, J., Barbero, L., 2003. Residence and redistribution of
803 REE, Y, Zr, Th and U during granulite-facies metamorphism: behaviour of accessory and major
804 phases in peraluminous granulites of central Spain. *Chemical Geology* 200, 293-323.

805 Watson, E.B., 1979. Apatite saturation in basic to intermediate magmas. *Geophysical Research Letters*
806 6, 937-940.

807 Watson, E.B., Green, T.H., 1981. Apatite/liquid partition coefficients for the rare earth elements and
808 strontium. *Earth and Planetary Science Letters* 56, 405-421.

809 Watson, E.B., Harrison, T.M., 1983. Zircon saturation revisited: temperature and composition effects in
810 a variety of crustal magma types. *Earth and Planetary Science Letters* 64, 295-304.

- Watson, E.B., Vicenzi, E.P., Rapp, R.P., 1989. Inclusion/host relations involving accessory minerals in high-grade metamorphic and anatectic rocks. *Contributions to Mineralogy and Petrology* 101, 220-231.
- Watt, G.R., Harley, S.L., 1993. Accessory phase controls on the geochemistry of crustal melts and restites produced during water-undersaturated partial melting. *Contributions to Mineralogy and Petrology* 114, 550-566.
- Webster, J.D., Piccoli, P.M., 2015. Magmatic apatite: a powerful, yet deceptive, mineral. *Elements* 11, 177-182.
- White, R.W., Pomroy, N.E., Powell, R., 2005. An in situ metatexite–diatexite transition in upper amphibolite facies rocks from Broken Hill, Australia. *Journal of Metamorphic Geology* 23, 579-602.
- White, R.W., Powell, R., 2002. Melt loss and the preservation of granulite facies mineral assemblages. *Journal of Metamorphic Geology* 20, 621-632.
- White, R.W., Powell, R., Holland, T.J.B., Johnson, T.E., Green, E.C.R., 2014a. New mineral activity–composition relations for thermodynamic calculations in metapelitic systems. *Journal of Metamorphic Geology* 32, 261-286.
- White, R.W., Powell, R., Johnson, T.E., 2014b. The effect of Mn on mineral stability in metapelites revisited: new a–x relations for manganese - bearing minerals. *Journal of Metamorphic Geology* 32, 809-828.
- Wolf, M.B., London, D., 1994. Apatite dissolution into peraluminous haplogranitic melts: an experimental study of solubilities and mechanisms. *Geochimica et Cosmochimica Acta* 58, 4127-4145.
- Wolf, M.B., London, D., 1995. Incongruent dissolution of REE-and Sr-rich apatite in peraluminous granitic liquids: Differential apatite, monazite, and xenotime solubilities during anatexis. *American Mineralogist* 80, 765-775.
- Yakymchuk, C., Brown, M., 2014a. Consequences of open-system melting in tectonics. *Journal of the Geological Society* 171, 21-40.
- Yakymchuk, C., Brown, M., 2014b. Behaviour of zircon and monazite during crustal melting. *Journal of the Geological Society* 171, 465–479.
- Yakymchuk, C., Brown, M., Clark, C., Korhonen, F.J., Piccoli, P.M., Siddoway, C.S., Taylor, R.J.M., Vervoort, J.D., 2015. Decoding polyphase migmatites using geochronology and phase equilibria modelling. *Journal of Metamorphic Geology* 33, 203-230.
- Yakymchuk, C., Clark, C., White, R.W., *in press*. Phase Relations, Reaction Sequences and Petrochronology in: Kohn, M.J., Lanari, P. (Eds.), *Petrochronology*, doi: 10.2138/rmg.2017.83.2.
- Yang, P., Rivers, T., The origin of Mn and Y annuli in garnet and the thermal dependence of P in garnet and Y in apatite in calc-pelite and pelite, Gagnon terrane, western Labrador. *Geological Materials Research* 4, 35p.
- Zeng, L., Asimow, P.D., Saleeby, J.B., 2005a. Coupling of anatectic reactions and dissolution of accessory phases and the Sr and Nd isotope systematics of anatectic melts from a metasedimentary source. *Geochimica et Cosmochimica Acta* 69, 3671-3682.
- Zeng, L., Saleeby, J.B., Asimow, P., 2005b. Nd isotope disequilibrium during crustal anatexis: A record from the Goat Ranch migmatite complex, southern Sierra Nevada batholith, California. *Geology* 33, 53-56.

Figure captions

856 **Figure 1.** P – T pseudosection for an average amphibolite-facies metapelite (modified from Yakymchuk
857 et al., *in press*). The dashed line is the solidus. The two isobaric heating paths are used to evaluate
858 closed-system versus open-system behaviour in Figures 6–9.

859
860 **Figure 2.** Contours for the proportion and composition of anatectic melt important for the modelling the
861 behaviour of apatite and monazite. (a) The proportion of anatectic melt in mol.% (approximately
862 equivalent to vol.%) in the metapelite in a closed system. (b) The aluminum saturation value of melt. (c)
863 the weight percentage of SiO_2 in melt. (d) the weight percentage of H_2O in melt. The dashed line is the
864 solidus.

865
866 **Figure 3.** Contours for saturation concentrations of melt in P_2O_5 (wt.%) using the apatite solubility
867 equations of (a) Harrison and Watson (1984): HW84, and (b) Wolf and London (1994): WL94. (c)
868 Contours for saturation concentrations of melt in LREE (ppm) using the solubility equation of Stepanov
869 et al. (2012) for monazite. The dashed line is the solidus.

870
871 **Figure 4.** Contours for the amount of apatite dissolution (%) in a closed system using the apatite
872 solubility equations of Harrison and Watson (1984): HW84, and Wolf and London (1994): WL94. The
873 assumed bulk rock concentration of LREE is 150 ppm and various concentrations of P_2O_5 are indicated.
874 The dashed line is the solidus.

875
876 **Figure 5.** Contours for the amount of monazite dissolution in a closed system comparing the difference
877 between excluding apatite from the system (e.g. Kelsey et al., 2008; Yakymchuk and Brown, 2014) and
878 including it. (a) Monazite dissolution contours (%) for the apatite-free scenario. (b) Monazite dissolution
879 contours (%) that include apatite using the Harrison and Watson (1984) apatite solubility equation. (c)

880 Monazite dissolution contours (%) that include apatite using the Wolf and London (1994) apatite
881 solubility equation. The dashed line is the solidus.

882

883 **Figure 6.** Mode–temperature diagrams illustrating the changes in modal proportions (mol.%
884 approximately equivalent to vol.%) along two isobaric heating paths from 650 to 950°C at 9 kbar (a,b)
885 and 6 kbar (c, d) for closed- and open-system scenarios. (e–h) The concentration of P₂O₅ (wt.%) in the
886 system during open- and closed-system scenarios for the two modelled isobaric heating paths. (i–j) The
887 concentration of LREE (ppm) in the system during open and closed-system scenarios for the two
888 modelled isobaric heating paths. WL94: apatite solubility equation of Wolf and London (1994). HW84:
889 apatite solubility equation of Harrison and Watson (1994). Melt loss (ML) events are indicated by the
890 vertical lines for the open-system scenarios.

891

892 **Figure 7.** Concentration of LREE and P₂O₅ in melt for the two isobaric heating paths in Fig. 1 using the
893 Wolf and London (1994) apatite solubility equation. Results for initial bulk P₂O₅ contents of 0.1, 0.3 and
894 0.5 wt.% for the open-system scenarios are shown. The concentrations of LREE and P₂O₅ in melt are
895 buffered to the saturation values when monazite and apatite are present. After apatite or monazite
896 exhaustion, the concentrations of LREE and P₂O₅ in melt are diluted during further partial melting.

897

898 **Figure 8.** (a–d) Amounts of apatite dissolution required to maintain phosphorus saturation of the melt
899 for closed- and open-system scenarios along two isobaric heating paths. In all cases, open-system
900 behaviour extends the stability of apatite to higher temperatures compared to the closed system. (e–h)
901 Amounts of monazite dissolution required to maintain LREE saturation of the melt for closed and open
902 system scenarios along two isobaric heating paths. In all cases, open-system behaviour extends the
903 stability of monazite to higher temperatures compared to the closed system. WL94: apatite solubility

904 equation of Wolf and London (1994). HW84: apatite solubility equation of Harrison and Watson (1994).
905 Melt loss (ML) events are indicated by the vertical lines.
906
907 **Figure 9.** Amount of monazite remaining relative to the amount present at the solidus for various molar
908 ratios of the LREE to the sum of all cations in monazite (X_{mnz}^{LREE}) for the open-system isobaric heating
909 paths at 9 kbar (a) and 6 kbar (b). The apatite solubility equation is from Wolf and London (1994). The
910 starting bulk concentration of P_2O_5 is 0.3 wt.%.
911
912 **Figure 10.** The relationship between temperature, time, grain size and the H_2O content of the melt for
913 apatite and monazite dissolution. (a) Apatite dissolution (modified from Rapp and Watson, 1986). (b)
914 Monazite dissolution (modified from Harrison and Watson, 1984).

915 **Table 1.** Bulk composition used in phase equilibria modelling (mol.%).

Figures	ML	H ₂ O	SiO ₂	Al ₂ O ₃	CaO	MgO	FeO	K ₂ O	Na ₂ O	TiO ₂	MnO	O
<i>Closed system – solidus saturated with H₂O at 9 kbar</i>												
1–6, 8		6.24	60.55	12.80	1.49	5.18	7.52	2.76	1.88	0.85	0.16	0.60
<i>Open system – Isobaric Heating at 9 kbar</i>												
	solidus	6.24	60.55	12.80	1.49	5.18	7.52	2.76	1.88	0.85	0.16	0.60
6–9	ML1	4.93	60.67	13.13	1.55	5.50	7.98	2.76	1.77	0.90	0.17	0.64
6–9	ML2	3.56	60.80	13.48	1.61	5.84	8.47	2.75	1.67	0.96	0.19	0.68
6–9	ML3	2.32	60.78	13.83	1.67	6.20	8.98	2.73	1.56	1.02	0.20	0.72
6–9	ML4	1.17	60.64	14.17	1.72	6.57	9.52	2.70	1.46	1.08	0.21	0.77
6–9	ML5	0.18	60.33	14.50	1.76	6.96	10.08	2.64	1.37	1.15	0.22	0.82
<i>Open system – Isobaric Heating at 6 kbar</i>												
6–9	solidus	5.15	61.23	12.94	1.50	5.24	7.61	2.79	1.90	0.86	0.17	0.61
6–9	ML1	4.10	61.10	13.27	1.57	5.56	8.07	2.77	1.82	0.91	0.18	0.65
6–9	ML2	3.16	60.84	13.60	1.64	5.90	8.55	2.74	1.73	0.97	0.19	0.69
6–9	ML3	2.27	60.48	13.93	1.71	6.25	9.05	2.69	1.65	1.03	0.20	0.73
6–9	ML4	1.40	60.05	14.27	1.78	6.63	9.59	2.62	1.59	1.10	0.21	0.78
6–9	ML5	0.81	59.36	14.58	1.84	7.01	10.13	2.53	1.53	1.16	0.23	0.83
6–9	ML6	0.38	58.51	14.88	1.89	7.41	10.69	2.41	1.49	1.24	0.24	0.88

916 *ML: melt loss event*

Figure 1

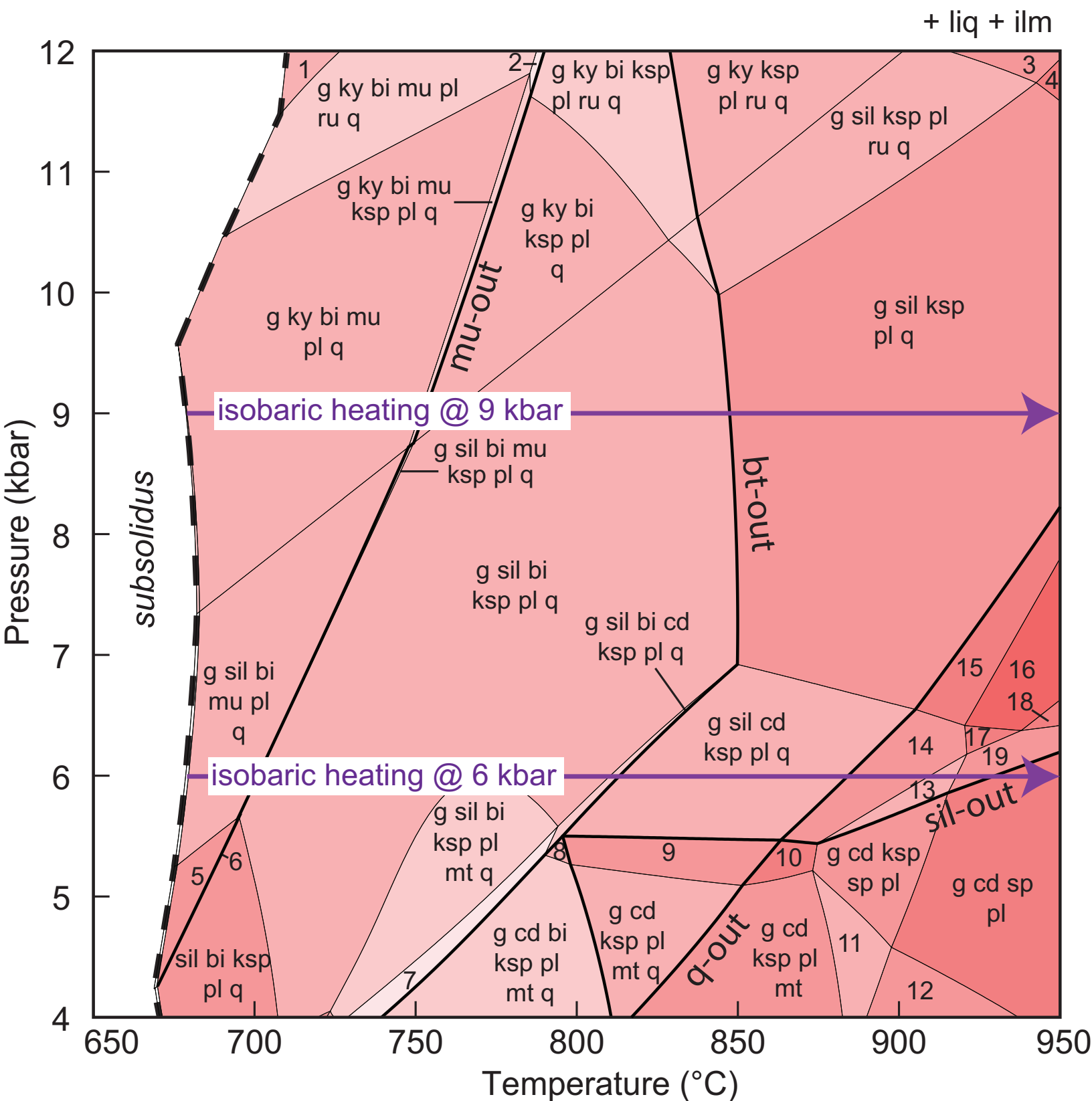
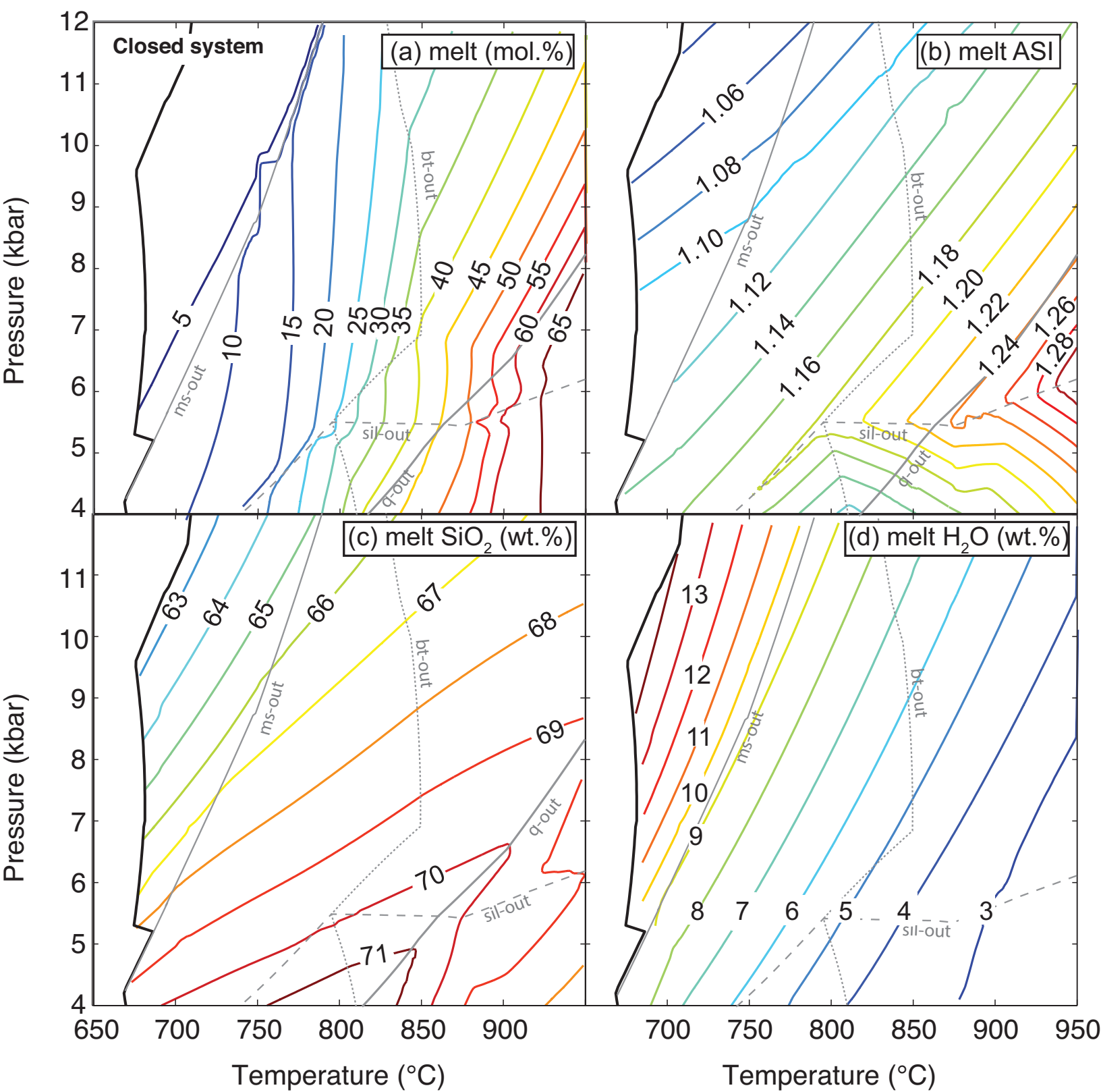
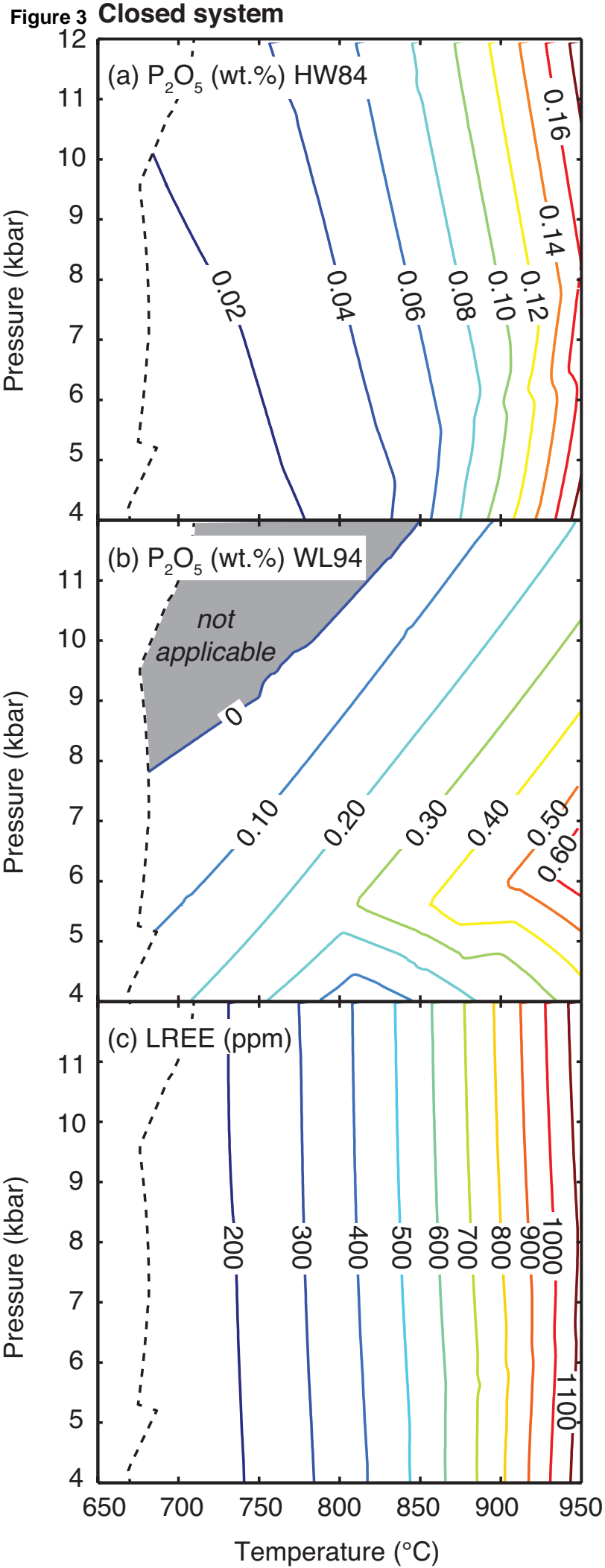


Figure 2





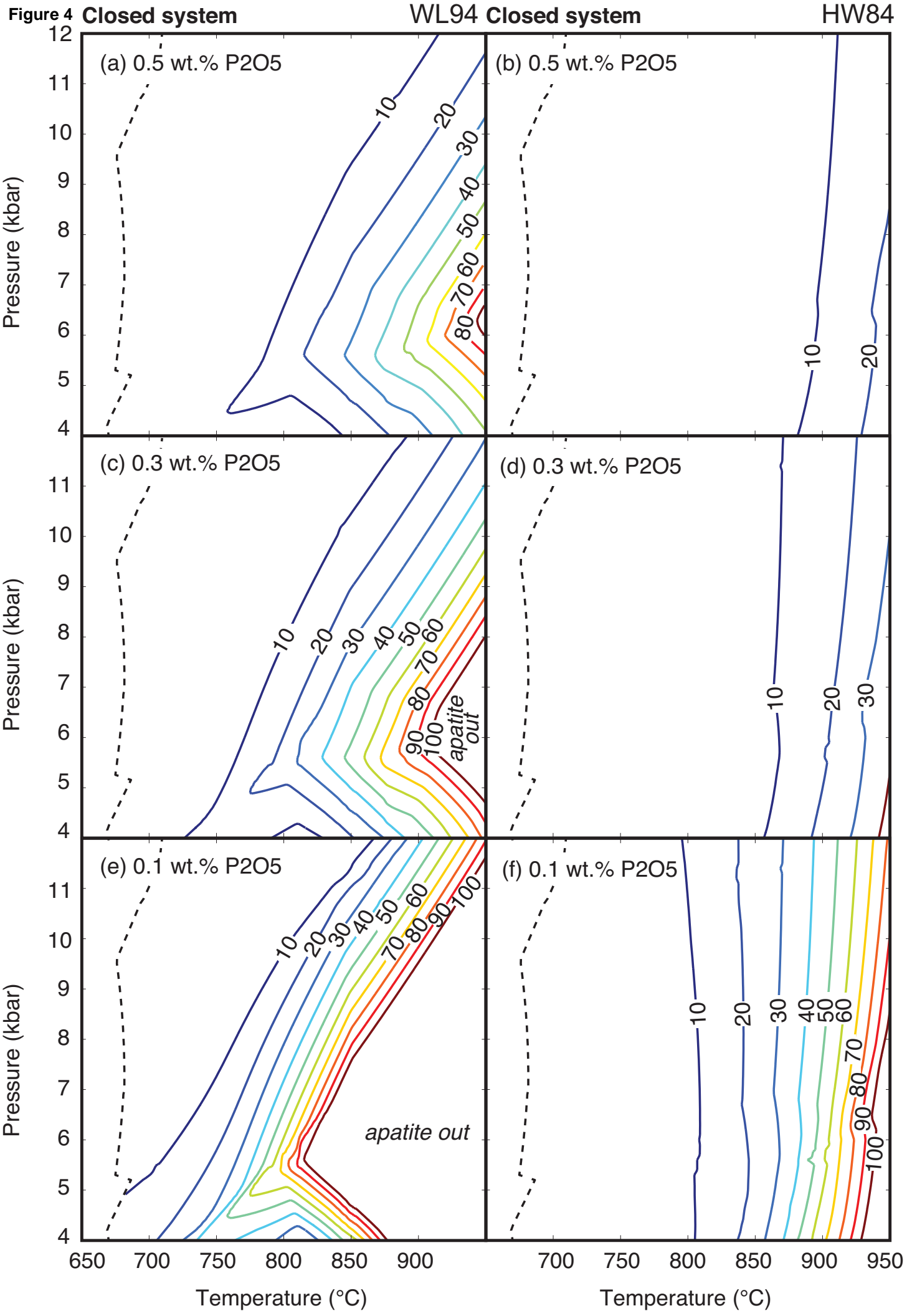
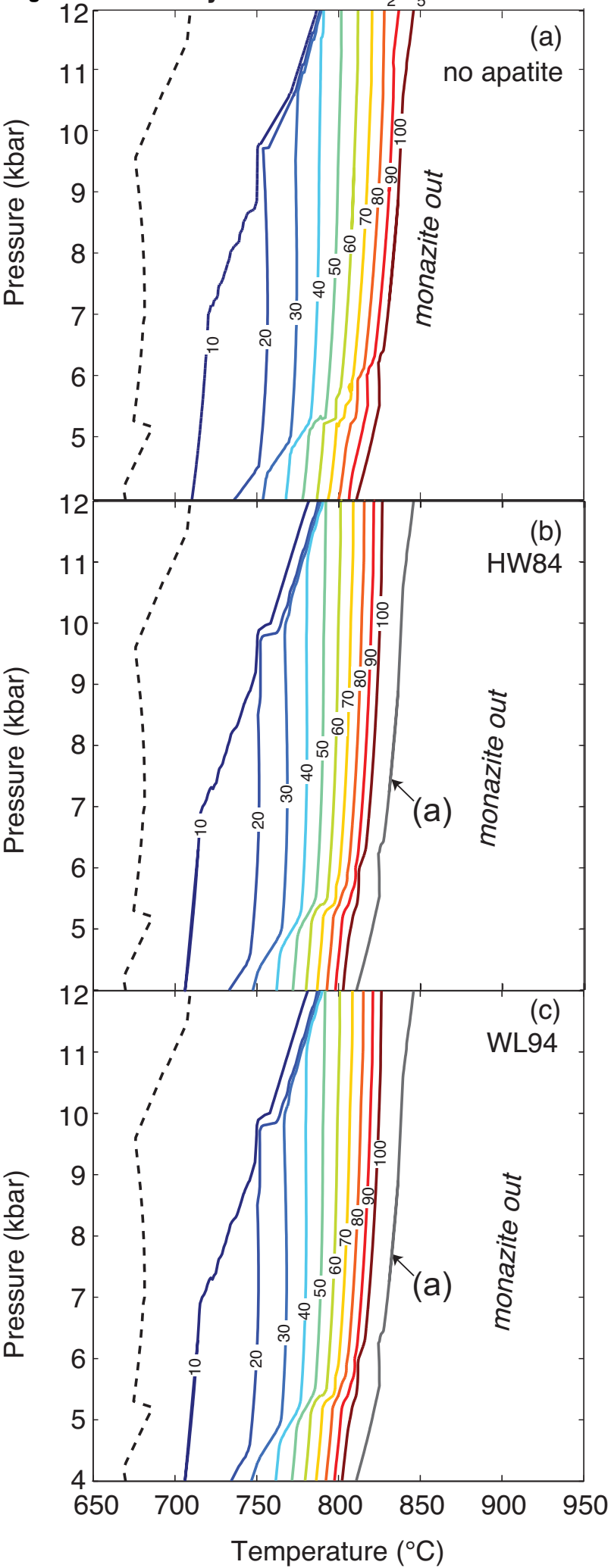


Figure 5 Closed system $P_2O_5 = 0.30$ wt. %



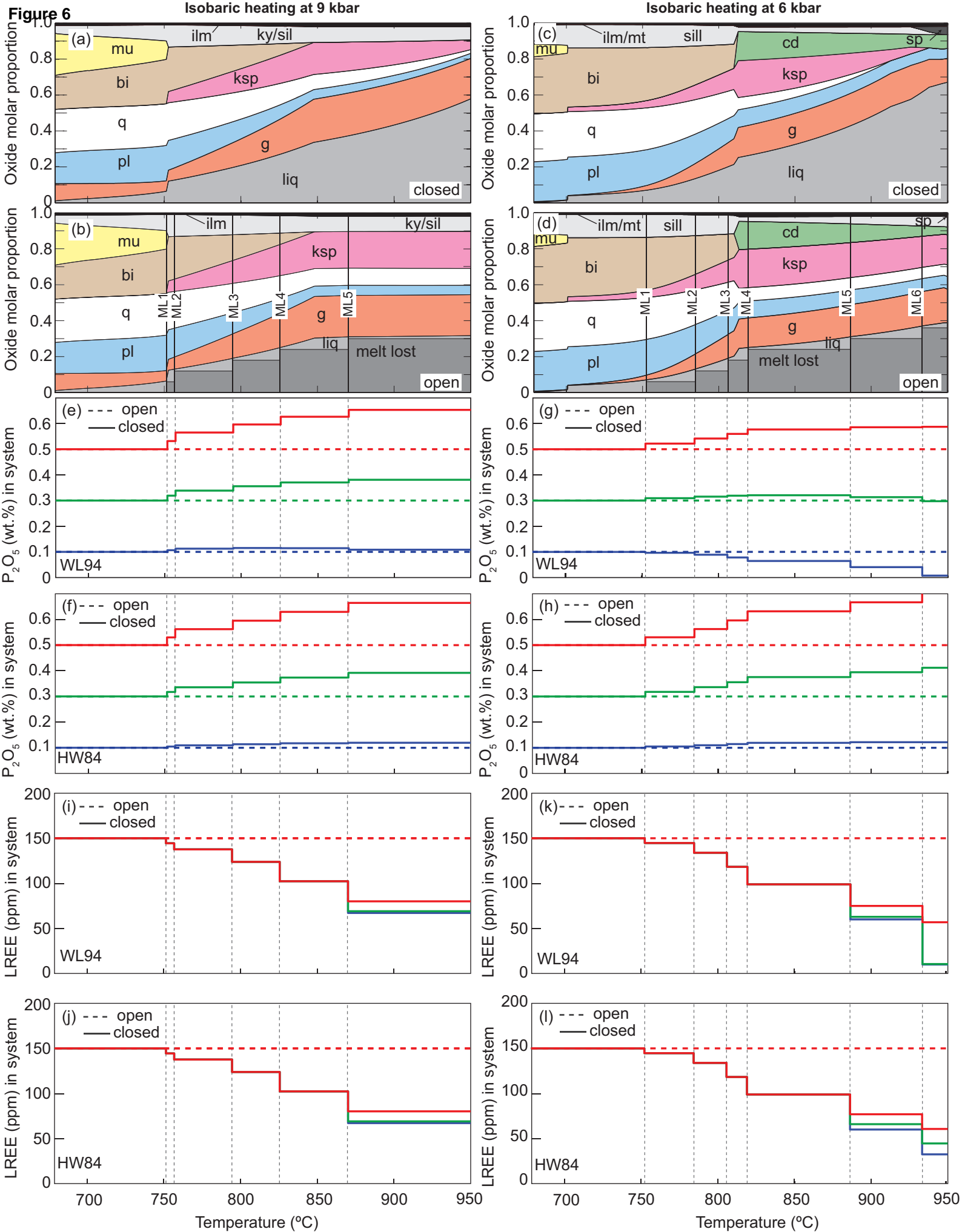


Figure 7

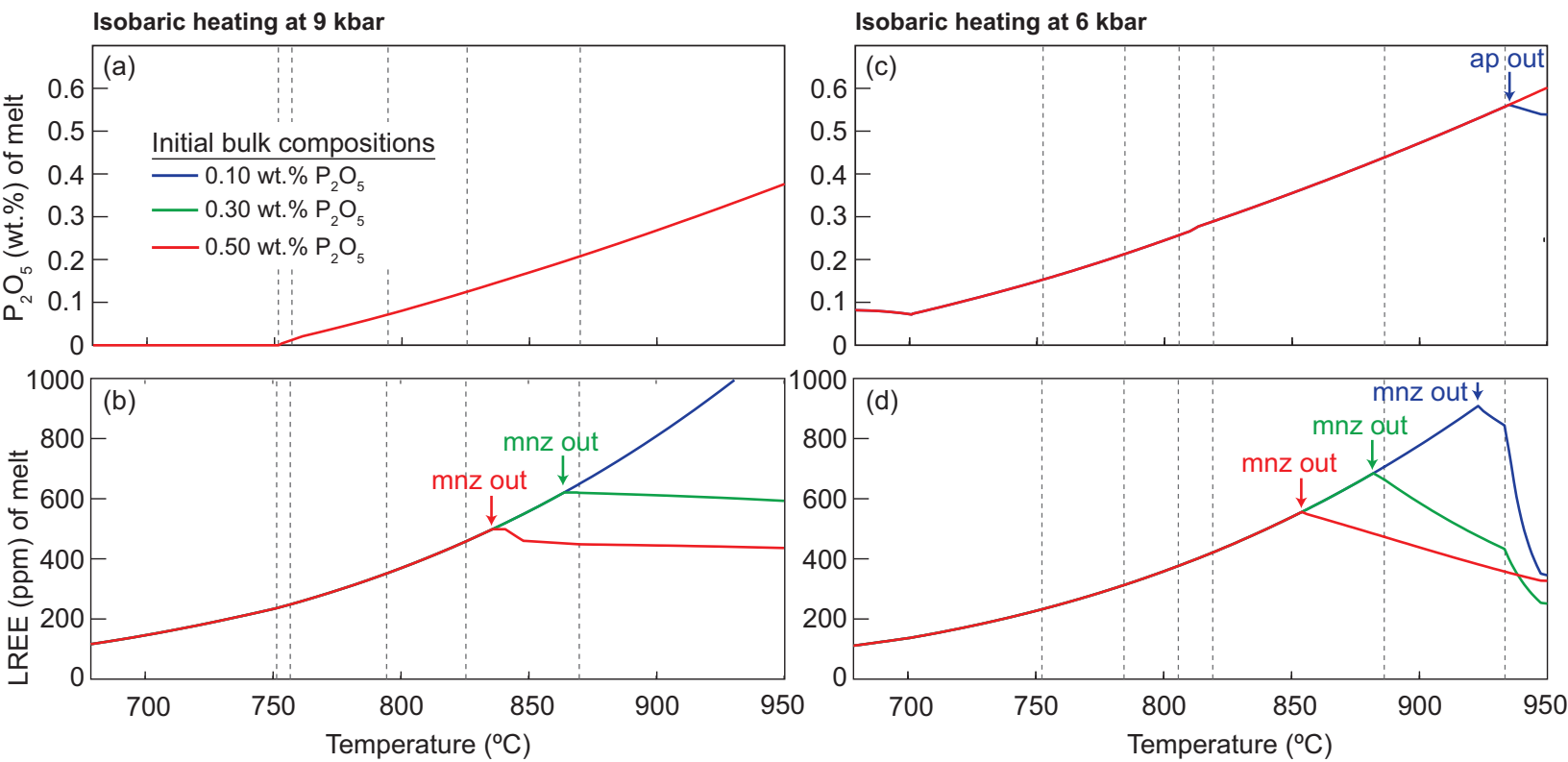


Figure 8

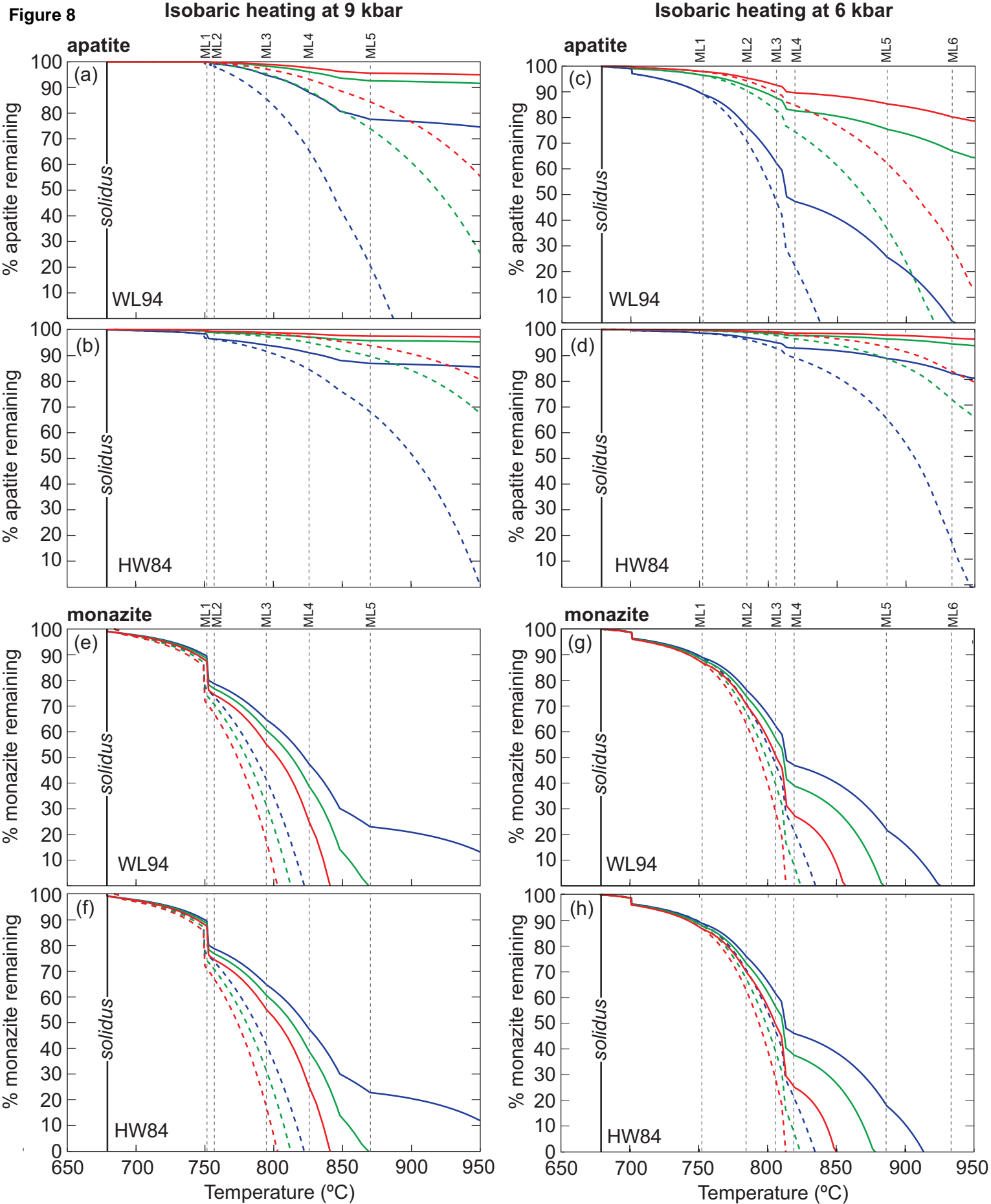


Figure 9

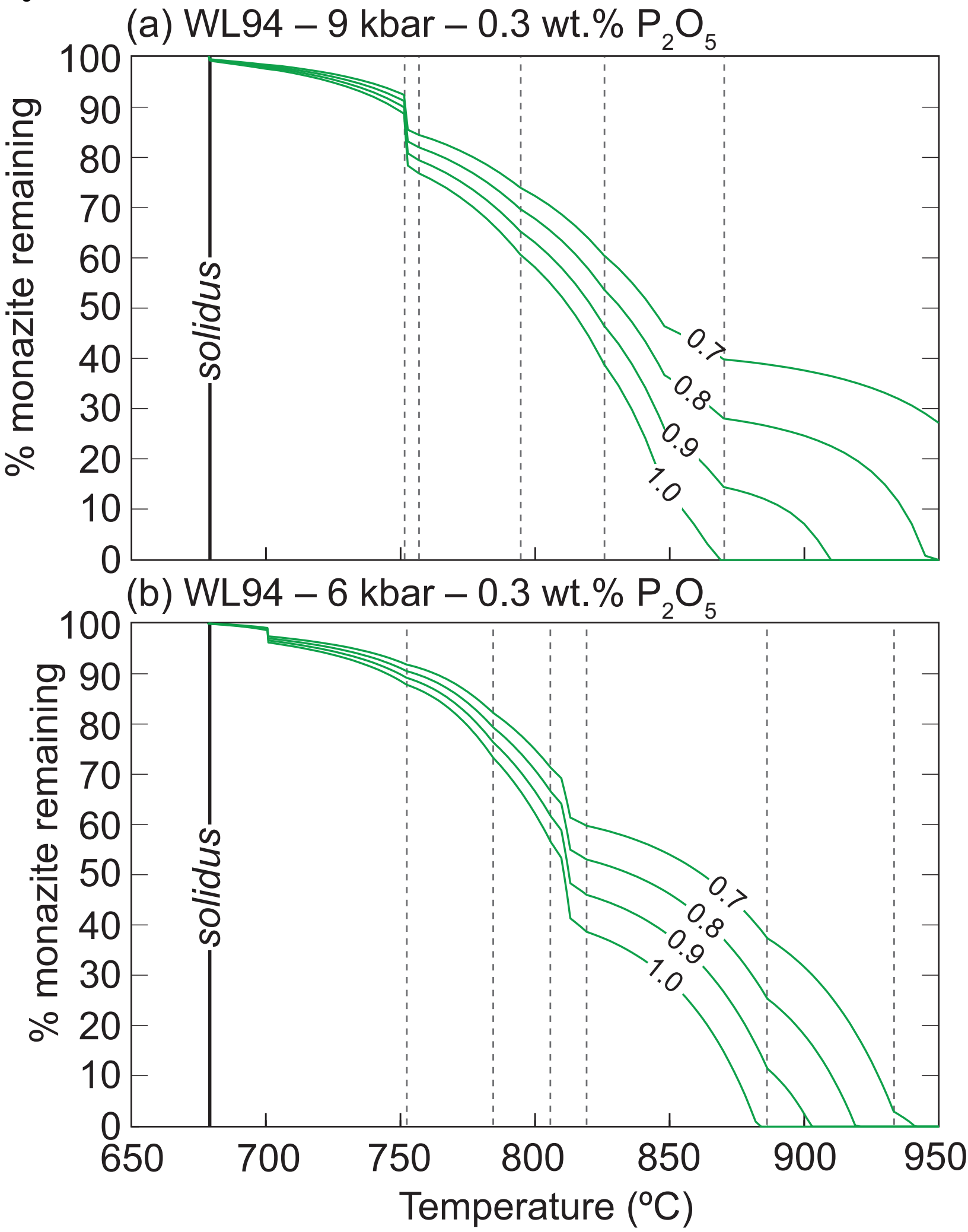


Figure 10

